GEOLOGY OF THE VALLE DE ARÁN (CENTRAL PYRENEES)

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CONTENTS

ABSTRACT

The results are given of field work undertaken in the central Pyrenean axial zone on low-grade metamorphic, highly folded Palaeozoic rocks.

Absence of determinable fossils is chiefly due to severe tectonisation, but occasionally also of non-deposition. Dating is entirely based on the presence of Silurian developed in a very persistent black shale facies.

The sandstones and shales of the Cambro-Ordovician show conglomerate horizons at different levels and a thickly developed limestone in the north and east.

The Silurian ampelitic slates are characterised by their high content of organic matter. Their rusty appearance in the field resulted from oxidation of the abundant pyrite.

The Devonian is developed in limestones and slates, but in the central part of the area sedimentation differs from the northern and southern parts of the axial zone; sandy deposits constitute the upper part of the Devonian sequence. Characteristic sedimentary structures together with the grading of part of the fine grained sandy deposits are considered evidence for residimentation by turbidity currents. Current directions measured from cross-laminations and convolutions indicate eastward directed transport of sediment.

Continuing emergence in the west during Carboniferous times resulted in erosion of part of the Devonian sequence, followed by rapid sedimentation of greywackes in a paralic environment.

Remnants of Triassic and late Miocene deposits are preserved north of the Maladeta granodiorite, probably as result of longitudinal faulting.

Cleavage-type folding took place during the Hercynian orogeny. A marked disharmony in folding between Devonian and Cambro-Ordovician resulted from the plastic properties of the carbonaceous Silurian slates. Late Hercynian faults, mostly longitudinal, are in some proved instances reactivated during following orogenies.

Observations on cleavage characteristics points to close relationship with and dependence on the lithology.

Lineations measured from intersection of bedding and cleavage run roughly parallel to fold axis and plunge.

Knicked cleavage, resulting from delatation, originated at the end of a late Hercynian uplift.

The dykes of the eastern part of the mapped area show coarse fracture cleavage, which developed after the normal cleavage. Some structural features are associated with the intrusion of large granitic masses.

Remnants of pre-glacial planation surfaces, presumably of post late-Miocene (Pliocene?) origin, were identified at three different levels.

Karst phenomena are frequent in the metamorphic Cambro-Ordovician and Devonian limestones. The large sink-holes are situated on or above the main planation surface, springs related to these sink-holes are found near the present river levels.

A geological map is provided showing lithostratigraphic subdivisions and five cross-sections.

INTRODUCTION

Most of the region surveyed — known as the "Valle de Arán" — forms part of the drainage area of the river Garona *), which has its sources here (Fig. 1). It contains the highest point of the Pyrenees, the granodiorite of the Maladeta s.s. **) with the Pico de Aneto (3,404 m).

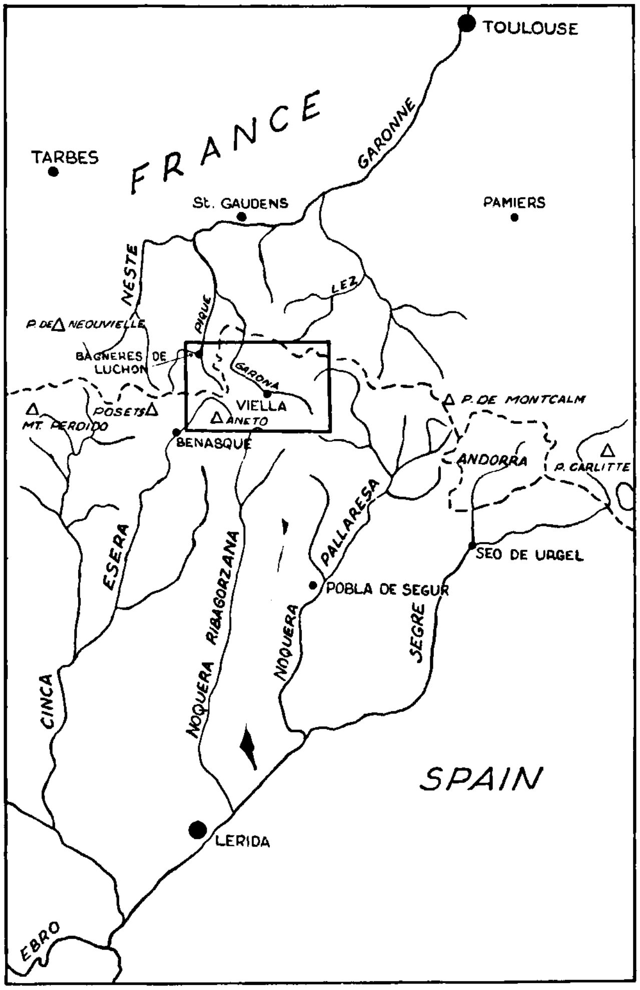


Fig. 1

*) Garona is the Spanish name for Garonne.
**) Maladeta s.l. the granodiorite batholith as a whole, Maladeta s.s. the highest western portion.

The lowest levels, of about 600 m, are found in the principal valleys of the Garona and Pique river systems.

The Valle de Arán upstreams of the Puente del Rey, a bridge over the river Garona, is Spanish territory, most of which is included in the province of Lérida; the south-western part of our map lies in the province of Huesca. The Valle de Arán is connected with the rest of the Spanish Pyrenees by the Puerto de la Bonaigua and by the tunnel between the Rio Negro and the valley of the Rio Noguera Ribagorzana. The French part of the geological map of the Central Pyrenees, sheet 4, is located in the departments of Haute Garonne and Ariège.

In 1951, geological surveys conducted by the University of Leiden under the guidance of Prof. L. U. De Sitter reached the north-eastern part of the Valle de Arán (Volker) and the vicinity of Bosost (Francken). In the following years other parts were studied; field training courses were twice held in this region during 1953 and 1954. The names of those who contributed to the work of mapping are given in an index on sheet 4.

The excellent topographical maps, scale 1:20,000, published by the "Institut Géographique National", and aerial photographs were used for the French side and the border regions.

The Spanish side was studied with the help of the Bosost, Isil, Benasque and Esterri de Aneu topographical sheets, scale 1:50,000, published by the "Instituto Geografico Catastral y de Estadistica de España". For short periods, aerial photographs covering parts of the area were available for consulting locally, and these provided valuable information.

Parts of the Valle de Arán were mapped in detail to scales of 1:5,000 (Las Bordas - Vilamos area), 1:10,000 and 1:12,500 (the Devonian of the Valle de Arán) and also to a scale of 1:25,000. These maps were enlargements of parts of the above-mentioned Spanish topographical sheets.

To facilitate map reading coördinates are given in the text. The coördinates of the Spanish part of the map are expressed in degrees and minutes, those of the French part are indicated in centigrades.

A great deal of information and help was obtained from Mr. Würth, chief engineer of "Productora Fuerzas Motorices", a company for the generation of hydro-electrical energy in the Valle de Arán.

Only incomplete results were obtained in the regions of extreme difficult access, such as the eastern slopes of the Rio Barados and the Rio Iñola. The geology of the Esera valley, south and west of the Plâ de Hospital, has only been mapped provisionally and will not be described here; it will be studied in greater detail in the near future and will be included on sheet 7. The metamorphic region of Bosost will not be dealt with either; a special treatise concerning this area by Dr. H. Zwart is in preparation and will be published in the "Leidse Geologische Mededelingen".

The climate north of the main watershed is influenced by the Atlantic Ocean: there is rain every month, especially during the winter period. South of it reigns a Mediterranean climate. The Pyrenees act as a climate barrier: abundant rains on the north side, an arid climate on the south side. Even in the region under discussion these differences in climate are perceptible. The frontier mountain range in the north-east constitutes the barrier for the relatively dry regions of the Marimaña, Plâ de Bérèt and Iñola. The rains in the western part of the region and on the high parts of the Maladeta s.l. are much more abundant. These differences in rainfall are seen in the flora.

Dense forests, consisting principally of conifers, are found especially on the northern slopes and in the valleys running from north to south.

In the west and north-west the timber-line is sharp; trees can be found up to about 1,700 m. In the east and south-east trees occasionally occur up to 2,100 m. On the north side of the Garona valley between Viella and Salardu, and in the drainage area of the river Iñola, trees are almost non-existent. Along the Rio Negro and the rivers further to the east, the lower parts of the woods consist of hazel, which sometimes entirely replaces the conifer.

Above the timber-line the mountains are covered by a hard, tough grass; only the highest parts of the mountains are barren.

Of the original fauna, the izard (the mountain goat or chamois) and the mountain fox still occur in the most isolated regions. The mountain bear and the wolf, which according to the many stories must have occurred frequently, seem to have completely vanished.

The population is chiefly concentrated in the main valleys of the rivers Garona and Pique. Urbanisation, which for a long time depopulated many mountain villages on the French side, is unknow on the Spanish side.

Agriculture is restricted to the main valleys, the pastures in many smaller valleys being reserved as hay-fields. Large flocks of sheep, horses, mules and cows graze during the summer season on the mountains above the timber-line.

The forests provide much timber, which is transported to the sawmills of Lès, Bosost and Mig-Aran. Several "carreteras forestales" have been constructed for this purpose. A great deal of timber is used in constructing the tunnels for the hydro-electric power plants. The high mountain ranges with their many glacial lakes in large basins on the one hand, and the deeply eroded valleys on the other hand, provide ideal conditions for the generation of hydro-electrical energy. Most of the imported labour is employed by "Productora Fuerzas Motorices".

Mining, which a short time ago still was of some importance, has totally disappeared as a means of subsistence; the mines have all been abandoned.

CHAPTER I

LITHOSTRATIGRAPHY

A. PREVIOUS AUTHORS

The Cambro-Ordovician sediments outcropping in the Northern Anticline were described as early as 1842, owing to the presence of valuable minerals occurring in a thick limestone of the upper part of the Cambro-Ordovician called the "calcaire métallifère" (Mussy, 1869). Stratigraphic tables were given by Leymerie (1881), Caralp (1888) and Roussel (1892, 1893, 1904).

The strongly metamorphic areas of Bagnères de Luchon, Bosost and Lés also attracted attention at an early date: Caralp (1888), Zirkel (1867), Leymerie (1858, 1862, 1870, 1881) and Roussel (1903). Bertrand (1907) gave a Palaeozoic stratigraphy based on petrographic facies and metamorphic grade. Influenced by the "nappe" theory, he attributed a Devonian age to the calcaire métallifère of the Northern Anticline. The stratigraphic table given by Dalloni (1930) is still much used. A good description of Palaeozoic sediments is given by Durand and Raguin (1943), but they were unaware of the many faults present in the Pic de Mauberme area (484/55). This may account for the many outwedging limestones and apophyses which appear on their cross-sections.

Since 1947 Destombes has published some detailed sections of Ordovician sediments observed in tunnels. A conglomerate horizon found in a tunnel in the "Hautes Pyrénées" (1949) about 130 m below the Silurian concurs very well with our observations. A detailed description of the Bentaillou region (485/58) mentioned a conglomerate horizon underlying the calcaire métallifère (1958). A thin limestone (our sandwich limestone) is correlated by Destombes with the calcaire métallifère, but these two limestones are of different stratigraphic levels. This author neglected probably the important fault between the calcaire métallifère and the Silurian, a fault already mentioned by Caralp.

Destombes and Raguin (1953) published a stratigraphic section of Ordovician (480 m) and Cambrian (1,020 m) sediments from the Lac d'Oo area in the Central Anticline, a few kilometres west of our map. The structural style of this region suggests that many repetitions are present in this section. The upper part, of more than 1,150 m, is given in Fig. 2, section 1, for correlation purposes with section 2 of the Port de Venasque area (462/44).

The Cambro-Ordovician sediments of the Marimaña area are described as Silurian by Caralp (1888) and Roussel (1904). Their opinion was not shared by Dalloni (1930), who mentioned Silurian (schistes carburés) near Tredós (485/45), the Puerto de la Bonaigua (490/42°.40′) and north of the Marimaña (4°.41′/41°.42′). The surrounding limestone was classified as Devonian. Schmidt (1931) mentioned conglomerates and quartzites from the Puerto de la Bonaigua which he thought represented the Caradocian. Snoep (1955) studied in detail the Plâ de Bérèt (4°.39′/42°.42′) and the surrounding areas. The black slates of the upper part of the Cambro-Ordovician north of the Marimaña were classified as Siluro-Devonian because of the absence of characteristic Silurian.

The basal limestone of the Devonian has been described as a "calcaire rubané" by Leymerie (1881) and Caralp (1888) from Casarilh (456/56) and from the "route forestière" to Superbagnères (455/55). Regarding the overlying slates as a second unit of carbonaceous Silurian, Caralp placed the basal limestone in the Middle-Silurian. This might explain why Destombes

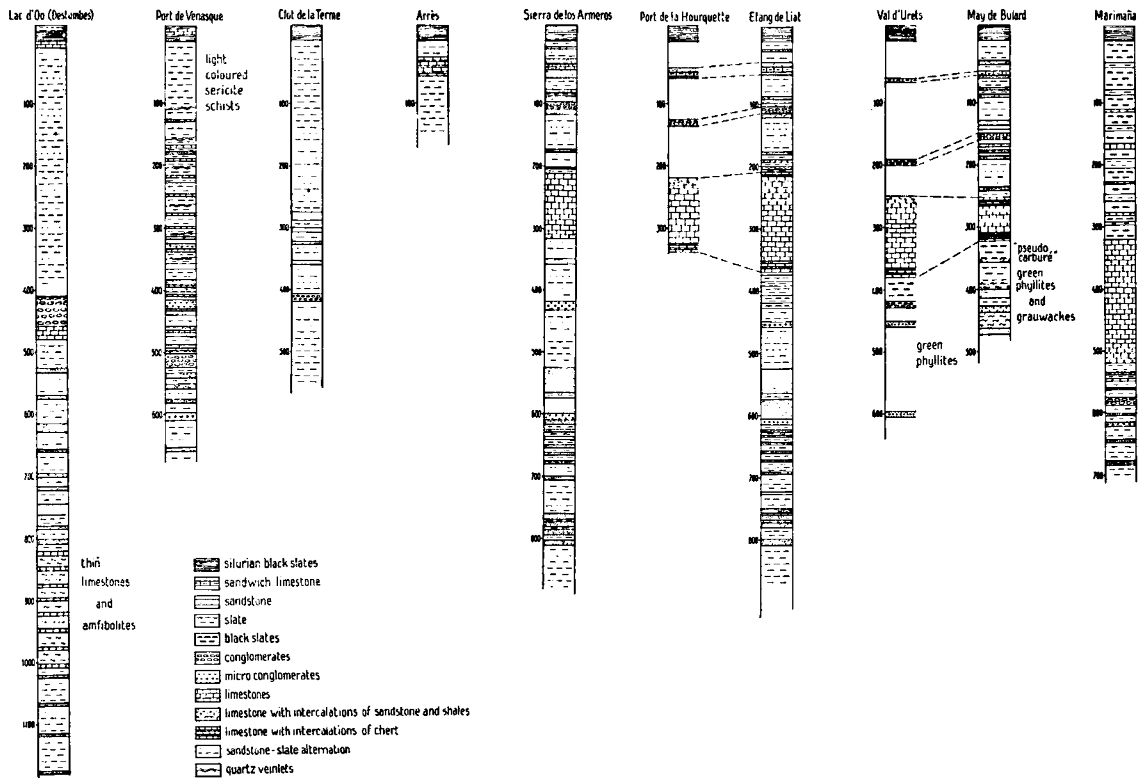


Fig. 2. Simplified lithostratigraphic sections of the Cambro-Ordovician

(1952), in this area, recognized a slate horizon as the base of the Devonian. Previous authors often mentioned two carbonaceous horizons belonging to the Silurian, a misinterpretation easily accounted for in these strongly-folded sediments (Caralp 1888).

The sandstones of the Las Bordas series have already been mentioned by Caralp (1888), Roussel (1893, 1903, 1904), Bertrand (1907) and Carez (1903—1909), who variously classified these deposits as Silurian (Caralp), Permo-Carboniferous and Silurian (Roussel), Carboniferous, Silurian, Cambrian (Bertrand), and Carboniferous (Carez). Dalloni (1930) and Schmidt (1931) held to the opinion of previous authors, Dalloni classified the sandstones as Lower-Silurian and Carboniferous, and Schmidt as Llandellian.

Calembert (1951) published a map showing the sandstones east of Begos (468/49) as Carboniferous.

Occurrences of Carboniferous sediments are mentioned by Gourdon; plant fragments found were determined by Zeiller (1886) as fragments of *Calamites, Lepidodendron, Sigillaria* and *Halonia*. Although they suggest merely a Carboniferous age, Dalloni later inferred a Westphalian age. In view of the abundance of plant remains, it seems strange that parts of the Carboniferous have occasionally been described or mapped as Silurian (Casteret 1931, de Lizaur y Roldan, 1951).

B. INTRODUCTION

The slightly metamorphic sedimentary rocks studied in this region are of Palaeozoic age, with the exception of very restricted Triassic, Tertiary and Quaternary deposits. It is difficult to apply a time-stratigraphical classification owing to a complete absence of index fossils in the central axial zone in contrast to the northern and southern borders of the axial zone, where marine fossils are present. The absence of marine fossils in the central part cannot be accounted for by non-deposition only (the main cause being complete destruction by strong tectonization), although this is probably the case for the upper part of the Devonian and for the Carboniferous.

Correlation has to be based on purely lithological characteristics and dating is possible only to a very limited extent. The latter is entirely based on the widely-distributed characteristic "black shale" formation dated by graptolites as Silurian. Formations older than this characteristic Silurian are classified as Cambro-Ordovician, and younger formations as Devonian and Carboniferous. The latter contain a flora which has been determined by Zeiller (1886).

The Cambro-Ordovician crops out in culminations called the Northern Anticline, Bosost Dome, Central Anticline and Marimaña Dome and will be described in that order. The region south of the Maladeta granodiorite will be mentioned together with the description of the Central Anticline.

It is impossible to fix the Silurian-Devonian boundary accurately, but in order to facilitate mapping it has been assumed that the Devonian begins with a limestone. This basal limestone usually rests directly on the carbonaceous slates, but where characteristic Silurian is overlain by non-carbonaceous slates they are included in the Silurian. As will be pointed out later, these non-carbonaceous slates still have some properties of the typical "black shale" facies.

A purely lithological subdivision has been made of the Devonian, partly on the basis of lateral changes in lithology and partly after a careful structural study. The description will start with the central part followed by a brief comparison of the north-west and south-west occurrences, in order of deposition throughout.

The Carboniferous and younger formations are described as a whole.

C. CAMBRO-ORDOVICIAN

a. The Northern Anticline and the Bosost Dome

The description of the Cambro-Ordovician is based on personal observations on the Spanish side and the detailed field work by Volker (the French side), van der Heyden (the extreme north-east on the French side) and Kapel (the Toran valley) (475/57) (internal reports). Lithostratigraphic sections are given in Fig. 2.

The Cambro-Ordovician sediments between the rivers Pique and Garona (458/46, Fig. 2 section 3) consist of a monotonous series of dark-coloured slates, grey slates, quartzites and sandy slates. The characteristic conglomerates and the calcaire métallifère known further east are much less developed or absent. Only north-west of Bosost (465/55) do strongly-tectonized conglomerates occur, consisting of flattened quartz pebbles 1—2 cm in diameter inbedded in sericite-schists rich in quartz. Calcareous sediments are absent.

West of the river Garona (sections 4—9) in the Urets area (483/57) the calcaire métallifère is situated between conglomerates; these conglomerates are found at about 150 m and 420 m below the Silurian. Correlation of the lowest conglomerate horizon with the Bosost conglomerate is most likely.

Dark-coloured slates, sandy slates and coarse-grained greenish-grey orthoquartzites and gravels occur near the village of Lès (467/57) and in the upper part of the Toran river (476/58). The lower part of the sequence (530—900 m in section 5) consists of dark, very fine-grained slates; the amount of quartz increases higher up in the formation, where coarse-grained calcareous quartzites occur, sometimes developed as micro-conglomerates. Banded sandstones ("schistes rubanés") consisting of thin sandstones alternating with slates are characteristic of this part of the sequence. The upper part (320—530 m in section 5) consists of sandstones, quartzites and dark-coloured sandy slates.

In the Sierra de Guarbes (473/54) a massive, sometimes marmorized, white-coloured limestone, the calcaire métallifère, overlies the dark, sandy slates. The basis of this limestone is black and shows intercalations of thin slates and chert. Slate and sandstone intercalations occur again towards the top. The calcaire métallifère thins towards the west, and is absent west and south of the river Garona. North of Arrès (567/51) isolated limestone bodies directly underlie the Silurian black slates (section 4). South of the village of La Bordeta (466/50) this limestone horizon, approximately 20 to 30 m thick, is exposed along the road. Its situation 50 m below the Silurian makes a correlation with the calcaire métallifère doubtful, although they look very much the same.

In the well-differentiated sediments in the central and eastern part (sections 5—9) a thin "sandwich" limestone occurs in the upper part (Fig. 3); this name, which describes very well its characteristic appearance, has been adopted from Volker.

The sandwich limestone, about 4 m thick, shows intercalations of chert, quartzites and siltstones up to 60 cm thick. The quartzitic layers decrease in thickness towards the top and become less regular; the top is formed by a calcareous quartzite. Where outcropping occurs, leaching has given this horizon a sponge-like surface.

The calcaire métallifère, a coarsely crystalline marble sometimes sugary in appearance, has its greatest extension east of Lake Liat (481/56) (section 7), slowly thinning towards the north (section 6), the south-west and west (section 5) and the east (section 9). Intercalations in the calcaire métallifère are sometimes abundant, for instance south of the Plâ de Tur (482/54) in the river Iñola, south and east of the Port d'Orle (490/55), north-east of the Etang d'Albe (481/59) and near the current Tartero (483/57).

In the eastern part the black-coloured base of the limestone contains abundant pyrite crystals. Dolomitization in irregular-shaped, light brown-coloured bodies, which obscure the bedding or schistosity planes, occurs occasionally, for instance near the Port de la Hourquette (482/57).

The conglomerates consist of well-rounded quartz and quartzite pebbles 1 to 15 cm in diameter, embedded in a schistose, quartzitic matrix containing chlorite and sericite. The pebbles, sometimes very large, change laterally into much finer ones or in gravel, and finaly into a coarse-grained sandstone. They are often flattened and this deformation is sometimes so strong that they appear as sandy lenses in a schistose matrix (Fig. 4). The ratio of the longest to the shortest axis varies considerably, but is generally about 8:3 (Fig. 5).

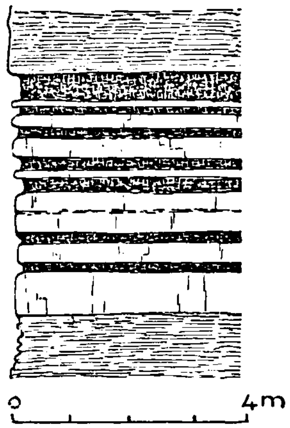


Fig. 3. Sandwich limestone from the upper part of the Cambro-Ordovician. After Volker

The matrix is sometimes calcareous, for instance north of Lake Liat, where the conglomerate seems to be higher in the formation than elsewhere. The conglomerates of the Pic de Past (section 8) (484/58) contain coarse limestone pebbles; they occur approximately 50 m above the calcaire métallifère.

A conglomerate also occurs below this limestone about 1,600 m north of the Port de la Hourquette in de Vallée d' Urets (485/57) (section 8) and is mentioned by Destombes in his recent paper (1958). It is much smaller in extent than the upper conglomerate horizon and is further only known south and east of the Marimaña area.

Between the "sandwich" limestone and the Silurian occur grey to dark-grey slates and occasionally thin quartzites or sandy slates. At about the same horizon as the conglomerate, grey and white banded sandstones overlie dark-coloured quartzite and sometimes calcareous slates.

Below the calcaire métallifère in the Urets valley (section 8 and 9) we find a formation comprising very dark-coloured slates, a black limestone band and black calcareous quartzites (the "pseudo carburé" or "série noir" of Destombes, 1958).

These dark series are underlain by black and white banded sandstones and light-grey or greenish-grey sericite-schists and quartzites. The "schistes

Fig. 4. Strongly tectonized Cambro-Ordovician conglomerates from the Mail de Bulard

Fig. 5. Cambro-Ordovician conglomerates from the Mail de Bulard

satinés" are very fine-grained, soft and talcose to the touch. The quartzites are often coarse-grained and contain micro-conglomerates 1 to 5 mm in diameter composed of clear quartz in a matrix rich in chlorite and sericite.

b. The Central Anticline

The Cambro-Ordovician rocks of this area consist of pelites which are phyllitic in character. Absence of good marker horizons and the intensive cleavage folding make it very difficult to estimate thicknesses. A stratigraphic section is given in Fig. 2 (section 2). The type of folding is illustrated in Fig. 57a.

The sediments of the upper part exposed in the Vallée du Port de Venasque (4°.20′/42°.42′) and in the upper reaches of the river Pique (4°.21′/42°.42′) are composed of grey or greenish-grey sericite-schists occa-

Fig. 6. Cambro-Ordovician conglomerates south of the Porte de Venasque

sionally alternating with dark-blue schists containing numerous quartz veinlets. In these schists we find occasional intercalations of grey-brown calcareous sandy bands. Lower in the formation the number of quartzites increases. Thin quartzites and sandy schists, often calcareous, alternate with limestones (not more than 1 cm thick). The lime is often leached from the sandy schists and quartzites. This has given them a honeycombed surface.

West of the Vallée du Port de Venasque the thin limestone intercalations in the schists and quartzites increase slowly in thickness to approximately 3 or 4 cm. The calcareous horizon is also thicker here.

The light-coloured non-calcareous quartzites are coarse-grained, sometimes micro-conglomeratic and have well-rounded, well-sorted and strongly flattened grains. The matrix consists of chlorite and sericite. An irregular conglomerate horizon sometimes occupies pockets in the quartzitic formation.

The quartz pebbles are very-well rounded, and are about 2 to 5 cm in diameter (Fig. 6).

South of the Pico de Escaleta (4°.22′/42°.42′) we find blue slates and sandy slates which often show a rusty brown colour near the Jueu fault; the light-coloured sericite-schists are absent here.

East of the river Jueu (4°.24′/42°.42′) only the upper part of the sequence, less than 400 m thick, is exposed. Soft, green-coloured sericite-schists are exposed east of the Rio Negro (4°.28′/42°.40′).

In the region of the Rio Bargadera (4°.37′/42°.40′) the Cambro-Ordovician sediments can hardly be distinguished from the overlying Silurian.

East of the Baños de Tredós (4°.37′/42°.39′) brown or grey-brown hornfels and spotted slates are exposed in the valley of the river Aiguamoix. These undoubtedly belong to the upper part of the Cambro-Ordovician.

South of the Maladeta s.s. in the Sierra Negra (4°.18′/42°.36′) the Cambro-Ordovician, probably occurring in the Southern Anticline, is obscured by Silurian debris. The absence of good exposures, however, makes it difficult to prove this.

c. *The Marimaña Dome*

The following description of the Marimaña region is based on the work of Snoep and van Alphen, but personal observations have also been used for the region of the Puerto de la Bonaigua (Fig. 2, section 10). Going from Salardu (4°.35′/42°.42′) to the Puerto de la Bonaigua one passes the sediments of section 10 from young to old. Black slates can be observed at the second hairpin bend on the road, followed by thick white marbles with abundant crinoid remains. This limestone, with its occasional irregularly-shaped dolomite bodies, looks the same as the calcaire métallifère, and occurs at nearly the same stratigraphical level. Blue-black slates with occasional limestones, coarse-grained schistose quartzites, light-grey serecite-schists, and occasional micro-conglomerates are found at the Puerto de la Bonaigua. A conglomerate horizon consisting of quartz and quartzitic pebbles up to 30 cm in diameter embedded in sericite-schists occurs south of the Puerto de la Bonaigua. North of the Puerto de la Bonaigua black slates are found underlying the calcaire métallifère. The lower part of the calcaire métallifère alternates with thin layers of slate, sandstone or chert with quartz-rich bands about ¼ to 5 cm thick (Fig. 7). The upper part is quite pure and contains occasional crinoid remains. The thickness of the calcaire métallifère very probably exceeds 200 m.

Hornfelses, found between the granite and the limestone, underlie this limestone. The original composition of these hornfelses (shale, fine- or coarse-grained sand particles and sometimes micro-conglomerates) is still visible in thin sections.

Small wedges of black slates representing younger series penetrate the limestone from the west. Towards the mouth of the river Malo (4°.38′/42°.42′) these pelitic sediments seem to be fairly thick; they contain occasional thin limestones and lighter-coloured slates. These slates, and the slates underlying the calcaire métallifère, have sometimes been classified by previous authors as Silurian.

At the south-west side of the river Ruda (4°.38′/42°.41′) a large slab of limestone overlies black, chiastolite-bearing slates. This limestone belongs either to the Devonian basal limestone or to the calcaire métallifère. North of this limestone slab another thin limestone with many intercalated chert layers is encountered. It does not resemble the calcaire métallifère of the Marimaña region, but it is still uncertain whether the limestones belong to the Devonian or to the Cambro-Ordovician.

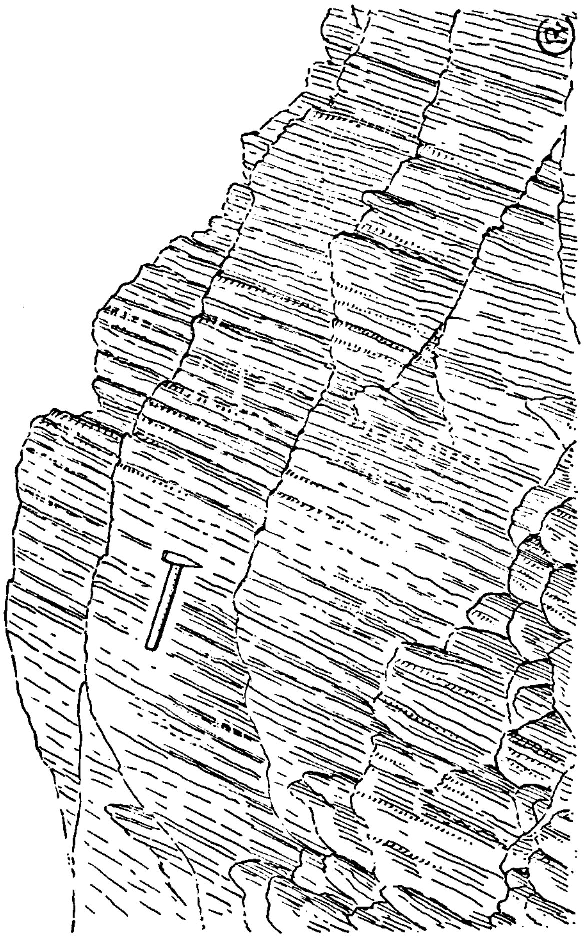


Fig. 7. Sandstone-limestone alternation at the base of the Cambro-Ordovician limestone south-west of the Marimaña. After photograph

North-west of the Marimaña the upper part of the calcaire métallifère grades into a calcareous slate. Black slates overlying the limestone cover a large area. They occasionally contain thin limestones with many shaly intercalations and small zones of light-coloured slates. Sandstones, or sandy slates, have been observed only in the north.

A gradual transition to the carbonaceous Silurian slates is difficult to find. The lack of good exposures on the Plâ de Bérèt is also a major difficulty. In the direction of the farm La Creu (4°.41′/42°.45′) thin zones of carbonaceous slates are found between the dark slates of the upper part of the Cambro-Ordovician, and Devonian limestones with slate and sandstone intercalations. A flexure is believed to separate the two units.

In the Iñola and Plâ de Bérèt areas (4°.37′/42°.43′) interpretation is severely handicapped by the great similarity between the upper Cambro-Ordovician and Devonian slates as well as by uncertain Silurian and poorly differentiated Devonian sediments.

D. SILURIAN

The Silurian occurs in a "black shale" facies as almost everywhere else in the Pyrenees. In spite of the great uniformity of the soft, very fine-grained black slates over great distances, it is sometimes hard to distinguish this horizon from the underlying Cambro-Ordovician slates. The boundary between the Silurian and the basal layers of the Devonian may also be difficult to discern in places where non-staining slates are intercalated between the Devonian basal limestone and the Silurian.

Another difficulty is encountered when extensive areas are covered with fine-grained detritus of the Silurian slates, as for instance in the Sierra Negra. It seems probable that they obscure the Cambro-Ordovician core of the Southern Anticline.

Neither the author nor previous workers have found graptolites in the area described, although they have been found frequently in adjacent regions such as the Pique valley and the Lez valley, where they allow a subdivision of the Silurian as is shown in Fig. 8.

Neither the Lower-Llandovery at the base nor the Upper-Ludlow at the top of the Silurian are characterized by graptolites, but there is no reason to conclude that these horizons are absent. There is no doubt that the non-fossiliferous "schistes carburés" of our region belong to the Silurian, and are therefore a valuable key-horizon.

The Silurian black slates of supposedly euxinic type have the following characteristics:

1. They are very fine grained and stain the fingers. The staining property of these slates is striking; where these slates are well-developed the landscape is coloured black.
2. The slates have a nearly normal iron percentage but a high sulphur content. Stratification or cleavage planes are often rusty-brown in colour. Small springs, often found in the upper part of the Silurian, stain the surrounding rock a rusty colour. An iron film is often seen on the surface of the water, and the water tastes strongly of sulphur. Slope breccias cemented by limonite are a common feature, especially in the Sierra Negra.
3. The slates show signs of strong deformation. Cleavage (or slickensiding) is well-developed but the planes are rarely regular. They have well-polished surfaces which have a metallic lustre, showing the same colours as seen in an iron or oil film on water.
4. Weathered slates sometimes have yellow-coloured spots. When struck with a hammer they give off an odour of hydrogen sulphide. Pyrite, mostly in well-developed crystals or as concretions, occurs abundantly. When strongly weathered the black slates lose their staining property, and may even become white.

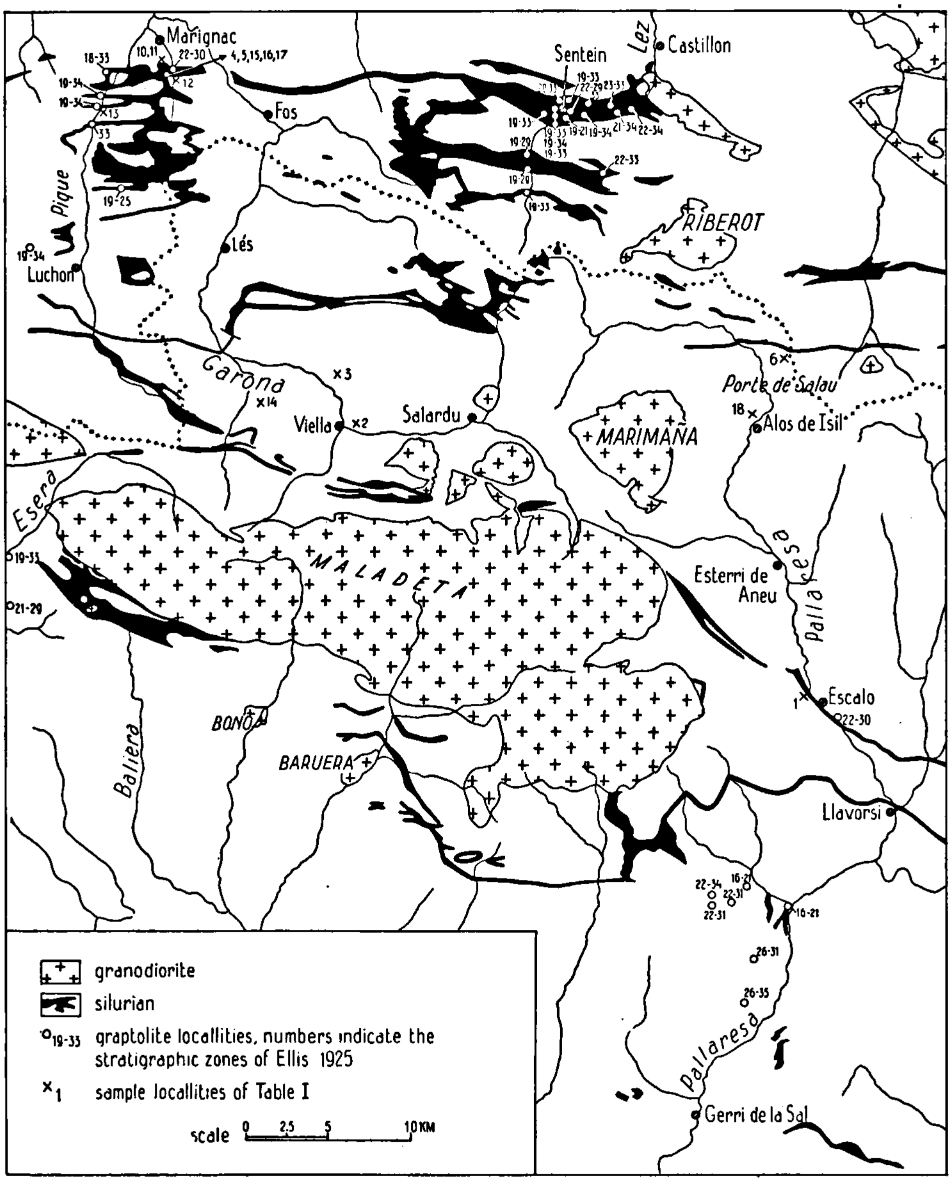

Marignac
Fos
Sentein
Lez
Castillon
Pique
Luchon
Lés
RIBEROT
Garona
Viella
Salardu
Porte de Salau
Alos de Isil
MARIMAÑA
Esera
MALADETA
Esterri de Aneu
Pallaresa
Escalo
BONO
BARUERA
Baliera
Llavorsi
Gerri de la Sal
granodiorite
silurian
graptolite locallities, numbers indicate the stratigraphic zones of Ellis 1925
sample locallities of Table I
scale 0 2,5 5 10KM

Fig. 8

5. In the metamorphic state the slates carry chiastolite. To the author's knowledge this mineral is only found in metamorphic Silurian and rarely in the metamorphic Carboniferous slates. Silica segregations and iron nodules (haematite?) are sometimes present in abundance.

Samples of Palaeozoic rocks have been taken for chemical analyses, and the localities are marked in Fig. 8. These analyses and those of Cap de Comme (1943) are given in Table I; they show great differences in carbon content.

According to Pettijohn (1957) a conversion from shale to slate does not appreciably change the gross chemical composition, but some change might be expected in the reduction of iron by organic matter and by the loss of CO_2.

The analyses of Silurian slates show a rather high carbon content, exept the samples 1 and 5, although they are both black, carbonaceous slates. This feature may be accounted for by the supposition that the slates are not uniformly carbonaceous; the carbon (or graphite) is often concentrated in layers between non-carbonaceous parts. This phenomenon has also been described by previous authors.

Cayeux (1935) mentioned a black Silurian limestone in which the graphite occurs in reticulate layers, enclosing varying amounts of lime. Cap de Comme (1943) was able to prove this for samples 12 and 13. Sample 12 from enriched layers yielded 18 %, 24 % and 24,5 % graphite.

Samples 1 and 5 are probably the non-carbonaceous parts split along the enriched layers. Although they easily stained the fingers on being touched, they actually have a very low carbon content. The percentage of quartz, iron and sulphur needs further discussion.

In these slightly metamorphic Palaeozoic slates quartz is often concentrated in veinlets. It is therefore not surprising to find great differences in quartz content in the samples.

Siliceous slates and silty slates have a relatively high silica percentage (Pettijohn). The finer the grain the lower the silica content and the higher the amount of alumina. Most clays contain proportionately large amounts of silt. This percentage decreases considerably in "black shales" of the euxinic type.

Silica in shales is present in clay minerals, as undecomposed detrital silicates and as free quartz of detrital or bio-chemical origin. The amount of free quartz is usually extremely low in black shales. The quartz percentage of the Silurian slates is considerably lower than that of ordinary slates.

The low sulphur content of the analyses is not consistent with the abundance of pyrite found in Silurian slates. This sulphur, supposedly of organic origin, is generally present in the form of pyrite. At the surface this pyrite is exposed to oxidation whereby in a humid climate limonite is formed. This limonite gives the rusty appearance characteristic of the Silurian. It has therefore often been thought that the Silurian slates must contain a higher iron percentage than is shown by the analyses.

The sulphuric acid formed during the oxidation process of the pyrite will combine with the lime — if present — to form gypsum, which is occasionally abundant in the Silurian. Most surface samples do not contain pyrite. In other cases the pyrite is concentrated in layers or in sometimes fist-sized concretions. Analyses of surface samples cannot be representative of the bulk composition of the rock as "black shales" of the euxinic type may contain up to 7 % of sulphur (Pettijohn).

TABLE I.

Chemical analyses of Palaeozoic slates from the Central Pyrenees.
Analyses given by Pettijohn are introduced for comparising.

	1 Silurian [2])	2 Devonian [4])	3 Devonian [4])	4 Silurian [4])	5 Silurian [4])	6 Ordovician [4])	7 Precambrium [3])	8 Palaeozoic [3])	9 Devonian [3])
SiO_2	40.40	40.60	57.90	?	?	52.20	58.03	60.15	60.65
TiO_2	1.48	0.66	1.47	1.20	1.15	1.24	0.64	0.76	0.62
P_2O_5	0.38	—	—	—	—	—	0.16	0.15	0.18
Al_2O_3	33.52	10.90	23.70	23.80	18.90	25.47	15.00	16.45	11.62
Fe_2O_3	2.43	6.50	5.76	?	7.00	6.30	3.67	4.04	0.36
FeO	4.28	—	—	—	—	—	5.82	2.90	—
MnO	0.10	—	—	—	—	—	0.09	trace	0.04
MgO	1.32	1.03	1.64	1.64	2.00	3.43	1.64	2.32	1.90
CaO	1.42	29.50	1.51	0.18	0.20	1.00	0.26	1.41	1.44
Na_2O	1.90	0.57	0.60	0.64	1.75	0.38	3.52	1.01	0.60
K_2O	4.95	2.80	5.42	4.20	3.98	6.51	3.60	3.60	3.10
H_2O+	6.72	—	—	—	—	—	3.46	3.82	3.77
H_2O-	0.20	—	—	—	—	—	0.84	0.89	1.19
CO_2	—	—	—	—	—	—	0.03	1.46	1.65
S_2	0.39	trace	trace	trace	trace	trace	0.04	0.58	3.20
C	0.34	4.70	0.24	5.80	0.32	0.54	3.27	0.88	9.20
	99.83	97.26	98.24	—	—	97.08	100.07	100.42	99.52
	carbonaceous slate	carbonaceous calcareous slate	dark grey slate	carbonaceous siliceous slate	dark grey micaceous silty slate	dark grey slate	black slate	composite sample of 51 Palaeozoic shales	Ohio shale
	power station Escalo	Las Bordas/ Viella formations	Las Bordas formation. Upper part of graded unit				Dunn Creek slates Crystal Falls district		Logan County
	Rio Noguera Pallaresa	north of Betrén Escuñau	north of Mont	Marignac area	Marignac area	Porte de Salau	Michigan		Ohio

	10 Silurian [1])	11 Silurian [1])	12 Silurian [1])	13 Silurian [1])	14 Devonian [4])	15 Silurian [4])	16 Silurian [4])	17 Silurian [4])	18 Ordovician [4])
	hard carbonaceous siliceous slate	hard carbonaceous siliceous slate	carbonaceous slate with quartz veinlets	carbonaceous limestone	dark grey hard micaceous slate	carbonaceous slate	carbonaceous silty slate	carbonaceous slate	dark grey micaceaus slate
	south of Marignac	south of Marignac	south of Marignac	Marignac area Lège	Entecada formation north-west of Corbison	Marignac area	Marignac area	Marignac area	north of Alos de Isil
C	3.80	5.70	6.10	4.14	0.14	4.50	8.80	5.25	0.48

[1]) L. Cap de Comme 1943.
[2]) analyst: Dr C. M. de Sitter-Koomans.
[3]) F. J. Pettijohn 1957.
[4]) Royal Shell Laboratory, Delft.
All elements except S_2 and C are spectrochemically determined.

The alumina percentage of the Pyrenean samples is very high, as can be seen in the Silurian samples 1, 4 and 5 and in the Devonian and Ordovician samples 3 and 6. In ordinary shale the amount of alumina is about 15 % (Pettijohn). The high alumina percentage may partly be due to the very fine grain size of the samples, and the low free quartz percentage.

The Silurian samples are all rich in potash; Devonian and Ordovician slates also have a high percentage of potash. The average potash content of a Palaeozoic slate is 3,6 % and does not exceed 5 % (Pettijohn); the potash probably originated from organic matter, like the carbon and sulphur.

Summarizing, we may say that the Silurian slates are characterized chemically by high carbon, sulphur, potash, and alumina contents. The percentage of iron is nearly normal and the silica content low.

Volker suggested that finely-divided sericite stained by the graphite might add to the staining property of the Silurian slates. The black material, which rubs off when touched may be black-coloured sericite.

Limestones occur in the Silurian between the rivers Pique and Jueu either as black lenses or as black nodules. Irregular quartz veins are common, for instance in the Mail du Criq (462/54) along the Bosost fault, or downstream from the Güells (4°.24'/42°.41').

The upper part of the Silurian sometimes consists of coarser material, but always shows some of the specific properties mentioned above, even if the slates are non-staining. This is shown, for instance, north of the Collado de Mounjora (464/45), between the Rio Jueu and the Rio Barados (476/52).

Black slates showing a very regular cleavage are exposed in the cores of the sharply-folded anticlines north of the village of Arties (480/45). Knaap (internal report) included these in the Silurian but except for the presence of pyrite, these black slates do not have any of the characteristics of the Silurian. Therefore, and also for other reasons (see page 152) these slates have not been classified as Silurian.

No clear distinction from Ordovician and Devonian slates can be observed on the Plâ de Bérèt, as has already been mentioned in the previous chapter. There is no reason to suppose that the Silurian here is not developed in the same "black shale" facies as everywhere else in the Pyrenees. Lack of good exposures might be partly responsible for the difficulty of determining this facies here.

It is hardly possible to estimate the thickness of the Silurian deposits. Strong deformation has often caused a secondary thinning or thickening. The best estimate can be made in the syncline of the Mail du Criq, where the Silurian deposits seem only slightly disturbed. Here some 200 m are exposed, which certainly represents a maximum.

Silurian slates are often found preserved along faults, as for instance in the Liat region, between La Renclusa (4°.20'/42°.40') and the Forat de Aigualluts (4°.21'/42°.40') and south-east of the Lago Bargadera (4°.31'/42°.39').

Chiastolite-bearing Silurian slates occur in the Northern Anticline, especially in the south-west corner, between the Pique de la Bonaigua and in the Sierra Negra. The chiastolite crystals are regularly developed and up to 10 cm long.

E. DEVONIAN

A subdivision of the Devonian is shown in Fig. 9. A number indicates the lithostratigraphic group, the second letter a lithological unit; estimaed thicknesses are given below from top to bottom:

Viella slates and sandstones D_4		
	very fine-grained green slates and sandstones	30— 50 m
Las Bordas sandstone D_3		
D_{3d}	graded sandstones and slates	200— 250 m
D_{3c}	non-graded sandstones and slates	50— 75 m
D_{3b}	orthoquartzites	20— 30 m
D_{3a}	sediments of a "littoral facies"	30— 50 m
Entecada slates and limestone D_2		
D_{2b}	grey-green or purple slates and sandstones	50— 100 m
D_{2a}	blue-black slates with detrital limestones	50— 75 m
D_2	limestone	10— 30 m
D_2	dark blue slates with occasional limestones and thin sandstones	50— 250 m
Basal limestone D_1		
	limestones with intercalations of chert, sandstone and slate	40— 150 m
	Total	530—1060 m

a. *Basal limestone* (D_1)

The Devonian starts with a basal limestone characterized by intercalations of dark chert and fine-grained sandstones at the base up to 2 cm thick (Fig. 10). Slate intercalations occur in the upper part, attaining thicknesses which may vary from 10 to 100 cm.

When it is marmorized the Basal limestone is white, as is illustrated by such geographical names as Peña Blanca, Roca Blanca, Mail Blanc, Tuc Blanc, etc. The weathered surface of the non-marmorized limestone may also be white, but the colour of a fresh fracture surface is usually dark grey to grey-blue or even blue-black. They contain pyrite, sometimes as large crystals, but more often in the form of spots or points like a pigment.

The upper part is well-exposed between the rivers Pique and Jueu, where it also seems to reach its thickest development. Remnants of crinoids are occasionally found, for instance north of the Collado de Mounjora. Along the Bosost Dome the D_1 thins considerably and west of the Pico de Trona (464/51) it is sometimes hard to detect. East of the river Jueu, in the Basal limestone on top of the Central Anticline, the intercalations are replaced by spherical, yellow to orange-coloured spots. Its minimum thickness here is about 40 m. South of the Negro fault, south of the Pico de Mompius (4°.26′/42°.42′) and in the Esera region (4°.17′/42°.41′) the characteristic intercalations, although present, are less well-developed, but in the Sierra Negra region the Basal limestone shows again frequent intercalations. Dolo-

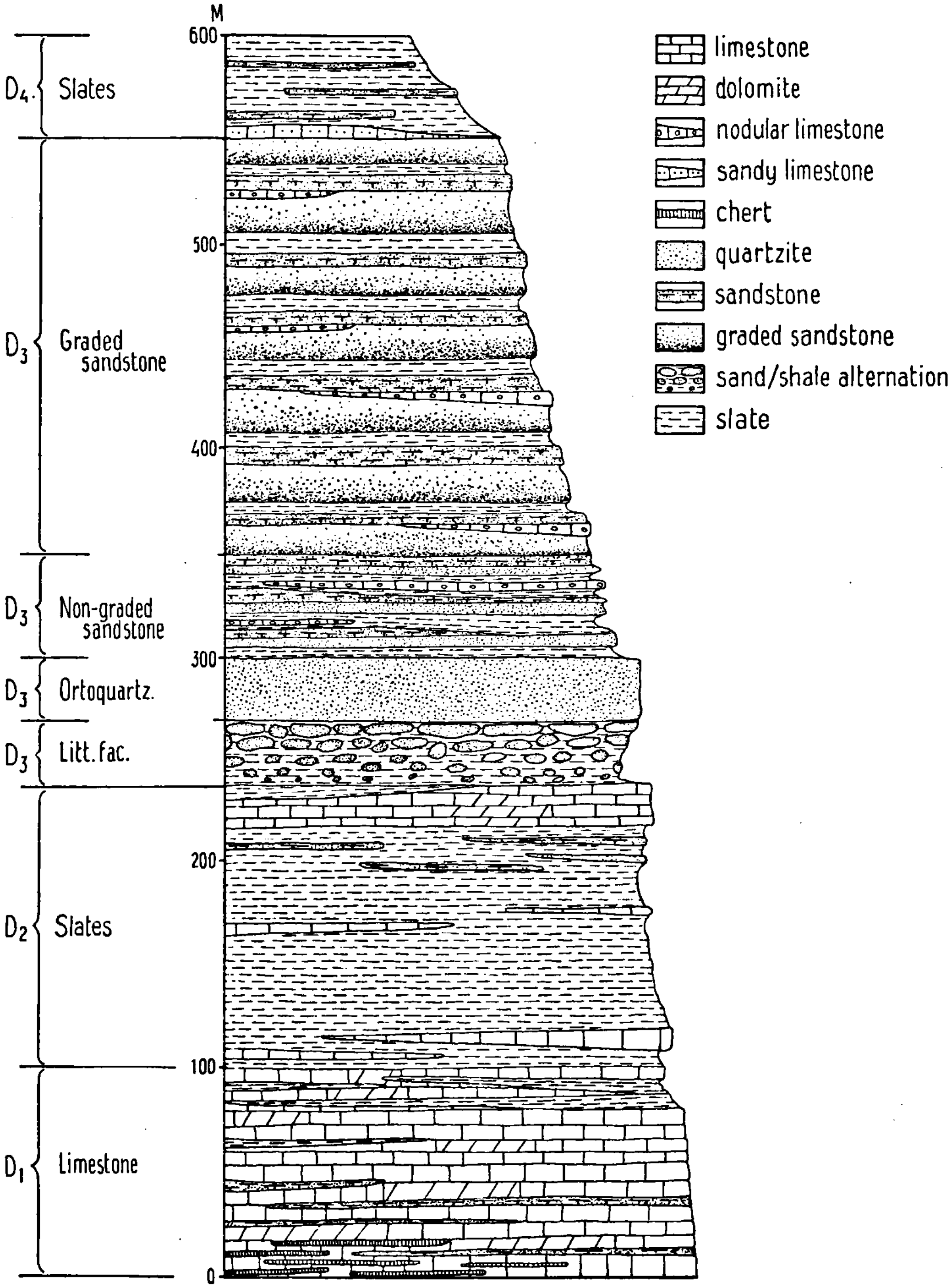


Fig. 9. Simplified lithostratigraphic table of the Devonian

mitisation is common; the dirty-brown coloured dolomite is distributed in irregular bodies, obliterating bedding and cleavage planes. West of the Pic de Aube (4°.26′/42°.41′) thin patches of very dark-coloured slate alternate with the limestone.

The thick, fairly pure limestone north of the Tuc de Media (4°.31′/42°.41′) is black and sometimes carbonaceous like the soil in this area. This indicates the presence of Silurian, but exposures of these carbonaceous slates were not observed. The granodiorite of Arties is surrounded by the Basal limestone as is proven by the occurence of Silurian south-east of Arties. The increase in hornfels next to the granite indicates probably the slaty upper part of the Basal limestone or even the Entecada slates. The age of the limestone pinched between the Tredós and Pruedo granodiorites and that of the limestone on the east side of the river (4°.36′/42°.40′) is still difficult to assess.

North of La Montañeta (4°.38′/42°.40′) a slab of limestone overlies thin,

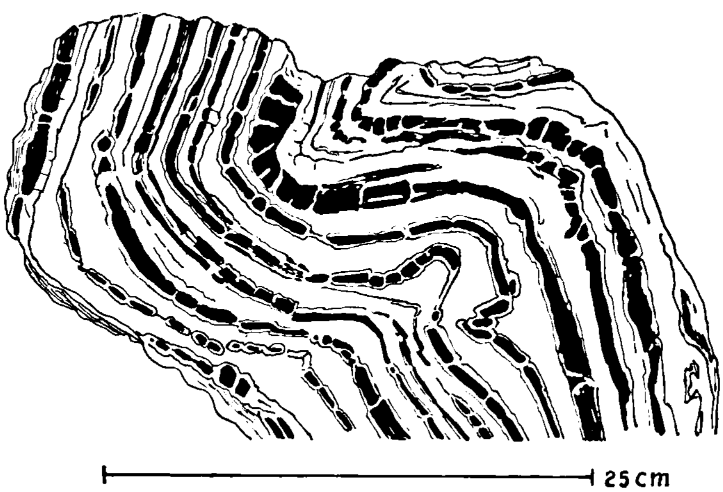


Fig. 10. Intercalating chert and calcareous chert in the lower part of the Basal limestone. Rio Jueu

non-carbonaceous black slates. These slates contain chiastolite, which might indicate a Devonian age for the overlying limestone. A thin limestone further north contains many thin sandstone and chert intercalations, characteristic of the Basal limestone, which do not resemble those of the calcaire métallifère of the Marimaña region. Thus it seems probable that the Tredós granodiorite is entirely surrounded by Devonian limestones.

The limestones which can be traced from Salardu towards the Puerto de Bérèt (4°.38′/42°.42′) are in the latter area accompanied by light-coloured calcareous slates and black slates. East of the Puerto de Bérèt they overlie the blue-black slates of the Upper-Ordovician, but the carbonaceous Silurian slates in between are missing.

The Basal limestone, normally with frequent intercalations, is found north and south of the abandoned village of Montgarri (not exposed on the map, east of 42°.45′.30″). Thick, marmorized limestones further west, however, are practically pure.

b. Entecada slates and limestone (D_2)

Dark-blue or blue-black homogeneous slates overlying the Basal limestone occur north-west of Bagnères de Luchon (457/55). Thin limestones are occasionally intercalated, usually at the base and near the top of this formation. The slates contain some pyrite in large crystals, but most of the pyrite is found as pigment. The wall of a tunnel built in this formation east of the river Jueu sometimes shows a red colour due to pyrite pigment.

The Entecada slates are well-developed north of the Pico de Entecada (466/47), and are named after this mountain. West of the village of Las Bordas (468/49) the upper part of the D_2 contains very thin sandstones and quartzites (5 to 20 mm) as can be seen in the bed of the river Garona.

In this region quartzites are also found alternating with the slates, but too few and too thin to be mapped separately. Further to the east the top of the D_2 formation is formed by a well-developed white limestone which is practically free of intercalations, underlying a thin layer (0 to 10 m) of slates. This limestone, particularly well-developed east of a line from Las Bordas to the Mompius, can be seen from Viella (474/46) on the eastern slopes of the Corbison (471/46), north of the river Garona, and east and west of the Rio Negro.

Dolomitisation is frequent; the dolomite occurs as irregular, light brown bodies in the limestone and obliterates completely bedding and cleavage. Bedding in the limestone is sometimes difficult to distinguish from cleavage or jointing.

Recrystallized crinoid remains are often found. Pyrite crystals in the Entecada limestone are sometimes found in thin parallel veins of quartz, a phenomenon also occasionally observed in the overlying and underlying slates.

Approximately 2 km west of the Las Bordas-Mompius ridge, sandstones occur in the top of the D_2 formation. Eastwards the sand content increases rapidly in importance and has been recognized as a very shallow-marine to littoral deposit. The Entecada slates directly north and north-west of Las Bordas are probably not as thick as in the Entecada region, along the Rio Barados and in the Mompius region. The Entecada slates are lacking altogether above the Basal limestone where it is overlain by the Carboniferous.

Near Las Bordas the slates have been converted into spotted slates. East of the Rio Barados in the surroundings of the Sierra de Betlán (475/51) the Entecada slates are slightly different. The lighter-coloured slates contain thin, micaceous sandstone laminae, increasing eastwards, and occasional quartzites. They can hardly be distinguished from the overlying graded sandstones. Here a limestone less than 10 m thick occurs near the top of the D_2.

In the Cabeza de Coll region (482/51) differentiation from the overlying graded or slightly graded sandstones is even more difficult than in the Sierra de Betlán area. South of the Monte Calvo (480/47) black slates underlying a thin limestone crop out in the centre of sharply-folded anticlines are darker than the normal Entecada slates.

Knaap (internal report) placed these slates in the Silurian because he thought the overlying limestone to represent the Basal limestone. The slates, however, are non-carbonaceous and do not show any of the other characteristics of the Silurian slates. On the other hand, the overlying limestone is rather thin (about 15 m) and pure, with only some rare and very thin shale intercalations, in contrast to the Basal limestone. The Entecada limestone is overlain by a thin layer of slates. This sequence is the same as mentioned from the Corbison region and there can be little doubt that the Monte Calvo sequence forms part of the Entecada formation.

The blue-black slates (D_{2a}) exposed east of the Areño (480/42°.44′) differ in many respects from the normal D_2 slates. They are slightly carbonaceous but do not show any other characteristic of the Silurian. They contain patches of detrital sandy limestone 1 to 10 cm thick, sometimes with well-developed cross-bedding.

In the middle of these slates, north of the Cabeza de Puch (482/47) a limestone which is thought to be the Basal limestone is exposed in an anticline. Hence these slates and detrital limestones are placed in the Entecada formation. The cross-bedding in the limestones is not found in the Candalias region (4°.37′/42°.44′) and towards the Plâ de Bérèt.

North-west of the Plâ de Bérèt and near the Pico de Parrós (4°.37′/42°.46′) the slates, greenish-grey, green or light purple have a thickness of 50 to 100 m (D_{2b}). They show properties similar to those in the Sierra de Betlán and Cabeza de Coll regions, as for example the presence of thin sandy laminae, but they also contain thin sandstones (10 to 20 cm thick). In the extreme north-east thin, blue-black-coloured slates overlie the Basal limestone.

Schists and hornfels which probably belong to the Entecada formation are found on the northern slope of the Montarto (4°.34′/42°.38′). Hornfels is found around the granodiorites east and west of the Rio Bargadera and west of the river Aiguamoix (4°36′/42°.41′), overlying the Basal limestone.

c. *Las Bordas Sandstones* (D_3)

In 1953 graded sandy deposits mapped as Devonian were observed near the village of Las Bordas during a field-training course. J. P. Snoep (1955) described sandy and graded deposits west of the Plâ de Bérèt and west of the village of Montgarri. All these sandy deposits are grouped under the name Las Bordas sandstones.

Although thin quartzites and sandstones also occur occasionally in the upper part of the Entecada slates, there is still a clear distinction between them and the Las Bordas sandstones; in the latter formation the sandy beds prevail over the slates.

In the Corbison area a lateral change in the lithology is observed. Small, very disturbed sand laminae occur frequently near the top of the Entecada slates, grading from west to east into pure quartzite. The relative proportions of sand and clay, always thoroughly mixed, may vary considerably (Figs. 11 and 12). These sediments, exposed mainly along the ridge of the Mompius and Corbison descending to Las Bordas, are thought to represent a very shallow-marine to littoral facies.

The quartzites into which they grade further east are fairly thick and can be followed along the eastern slope of the Corbison as far as the village of Gausach. Along the road from Las Bordas to Viella and Salardu, quartzites are also common, but the beds never exceed 1 m in thickness. The occurrences of very thick quartzites seem to be limited to the eastern slope of the Corbison. Towards the north and east they are overlain by an alternating series in which homogeneous quartzites and sandstones, sandy slates and slates with occasional thin limestones predominate. They are called "non-graded D_3" because graded sandstones have only occasionally been observed.

Grey or dark-grey graded sandstones are very well-developed in the Areño area and between the villages of Vilach and Mont (473/48), where they are mostly green or light-green. It often seems as if the green graded sandstones form the lower part of the graded deposits, but this is nowhere conclusively shown.

The grey sandy D_2 slates of the Cabeza de Coll pass gradually into the

Fig. 11. Strongly disturbed sandstone and shale laminae of the Las Bordas formation (D_{3a} littoral facies) south-east of Corbison

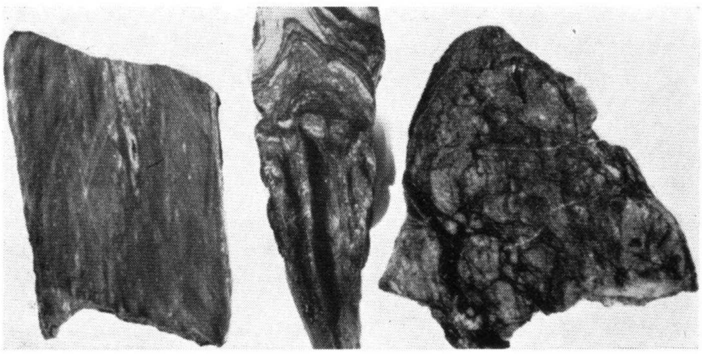


Fig. 12. Varying amounts of sand in samples of the Las Bordas formation (D_{3a} littoral facies) in the Corbison area

graded sandstones of the D_3 formation. Homogeneous sandstones and quartzites, such as are found north of a line from Salardu to the Monte Corbison are probably absent here, as is also the limestone of the Entecada formation.

North of the village of Garos (4°.32′/42°.42′) a nodular limestone about 50 cm thick has been found between the sandstones. The graded beds can also be fairly calcareous, but thin limestones are rare.

In the Cabeza de Portans area (4°.38′/42°.41′) and north-east of the Candalias graded sandstones are less well-developed. Further to the east sandy slates gradually replace the siltstones, sandstones and quartzites;

intercalations of thin limestones are common. In the massive, homogeneous sandstones of the Montgarri area grading is no more observed.

The sandy slates south of the Cabeza de Portans can hardly be mapped separately; the subdivision of the Devonian into sharply-divided lithological units is less pronounced here.

d. Viella slates and sandstones (D_4)

Soft, fine-grained, green slates with thin sandstones overlie the Las Bordas formation east of Viella, east and west of the Rio Negro and along the road to the south. South of the Monte Calvo these green series form

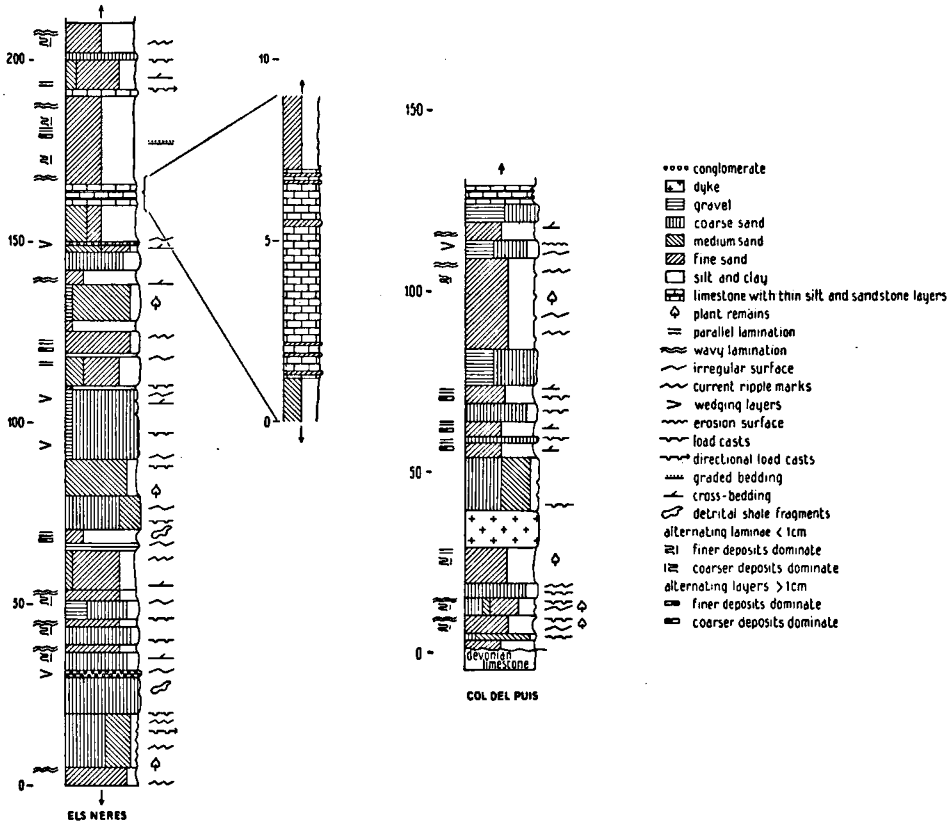


Fig. 13

as well the upper part of the lithological succession in this region. At the base of this formation there occurs usually a sandy, granular, or sometimes shaly limestone (less than 2 m thick) called "chalco-schists" by Snoep (1955). Greenish slates have been observed in the D_{2b} and Las Bordas sandstones. Some of them may correspond to the Viella green slates, but no lateral connection has been observed.

South of Viella the Viella slates and sandstones are probably some

50 m thick but thinner towards the east. West of the Puerto de Bérèt there remains only an alternation of thin green and greenish-grey slates, calcareous slates and limestones which cannot be subdivided further.

F. CARBONIFEROUS

The Carboniferous occurs in an east-west trending synclinorium north of the Maladeta s.l. The composition is shown in the sections of Fig. 13. The contact of the Carboniferous with the underlying Devonian is either faulted or unconformable, as shown by the complete absence of the Entecada, Las Bordas and Viella formations.

In places, thin slate layers probably belonging to the Carboniferous are

Fig. 14. Shale fragments in coarse-grained greywackes. Upper Rio Negro

found between the Carboniferous greywackes and the Devonian basal limestone. The greywackes are thick-bedded, coarse-grained and micaceous, occasionally grading into micro- or coarse conglomerates. Finer-grained sandstones, greywackes and sub-arkoses alternating with thin slates constitute a minority. In general the coarser the grain the lighter the colour, showing that the carbon content decreases with increasing grain size. The dark, finer-grained greywackes and slates are often carbonaceous and in metamorphic state contain abundant small chiastolite crystals.

The conglomerates consist mainly of well-rounded quartz pebbles forming irregular layers or lenses in the greywackes. Thick layers of micro-conglomerates mainly consisting of angular quartz occur abundantly.

Irregularly-shaped shale fragments (1 to 20 cm) are frequently found in the greywackes (Fig. 14). In different parts of the Carboniferous micro-folded, often dark-coloured limestone or sandy limestone alternate with thin, fine-grained sandstones or siltstones. These intercalations are especially frequent in the upper and lower parts of the limestone. The greywackes may be ferruginous; when weathered, stratification planes or joint planes are red or reddish-brown in colour. Shallow-water sedimentary structures mostly small in size occur abundantly in the slates between the coarse greywackes. Thin layers or irregularly-shaped pockets of carbonaceous slates are intercalated between the thick, coarse greywackes. Thicker layers of slates and siltstones, sometimes laminated, form only a minor part of the Carboniferous.

Abundant plant fragments are found, especially in the greywackes and sandstones; "in situ" origin of these fragments seems unlikely. Well-preserved prints of plant leaves have never been observed in the slates, although these are truly carbonaceous. The abundance of sedimentary structures and plant

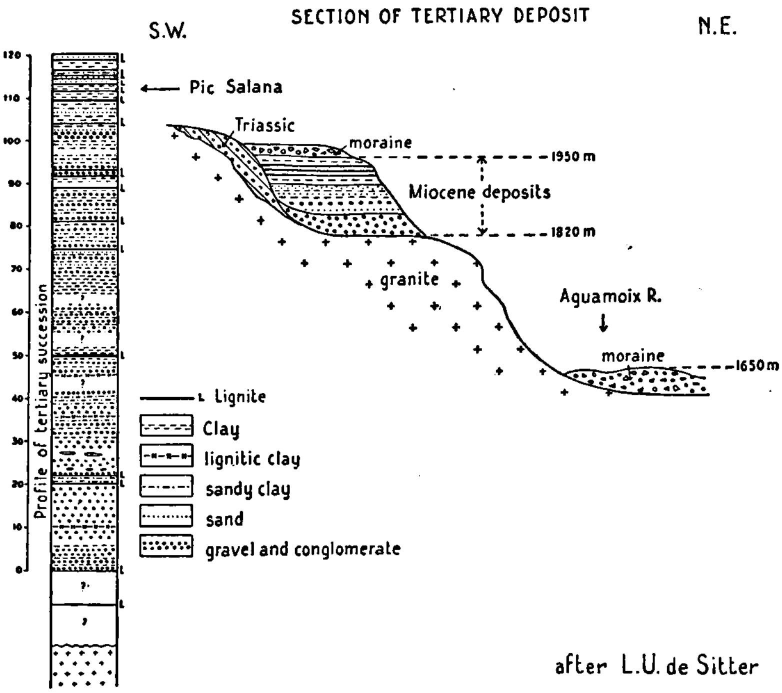


Fig. 15.

fragments indicates that the Carboniferous sediments are deposited in a paralic environment.

Towards the east the composition of the Carboniferous sediments changes slowly from psammitic to pelitic.

G. TRIAS

On the north-east flank of the Pico de Salana (4°.36'/42°.34') occur small outcrops of red-coloured sediments of typical Triassic appearance, mainly consisting of sandstones with occasional micro-conglomerates, grits and purple mudstones.

Good exposures are rare owing to the blanket of moraine material. The occurrence of Triassic sediments is, as far as known, unique in the centre of the axial zone of the Pyrenees.

H. TERTIARY

On the western bank of the river Aiguamoix (4°.36′/42°.40′) Volker in 1953 found lignites and conglomerates which he thought to be of Tertiary age. Later Miss S. Jelgersma analysed the pollen of these deposits, comparing them with the pollen associations from the well-dated Miocene (Vindobonian) deposits of Estavar in the Cerdaña region in the south-east Pyrenees. This study indicates a late Miocene age for the unique Aiguamoix occurrences (S. Jelgersma, 1958).

The horizontal, well-bedded Miocene deposits are exposed in some small gullies and surprisingly have not been eroded during the glacial period. The horizontal extension of the Miocene is not well-known, the greater part being covered by thick glacial deposits.

The Miocene starts with conglomerates consisting of very well-rounded pebbles mainly derived from sandstone (Trias), limestone, shale (Devonian?), hornfels, and rarely granite. The pebbles are embedded in coarse sand, mainly composed of the same components as described above. The conglomerates are overlain by coarse sand with occasional small bands of dark-brown lignite and pure, light-grey clay. Towards the top the amount of clay increases (Fig. 15). The sequence contains altogether 22 bands of brown, amorphous and brittle lignite varying in thickness from 4 to 62 cm.

In the excursion guide for the V International Quarternary Congress (Inqua, 1957) Sole Sabaris wrote that the morphology of this area and the facies relationships of the deposited sediments suggested an early-interglacial deposition of the lignite-bearing sediments. The pollen analyses, however, strongly contradicts this opinion.

CHAPTER II

STRUCTURAL GEOLOGY

A. GENERAL

For a long time the Pyrenees have greatly interested structural geologists. Although many folding theories have been applied to this mountain chain, little was known about the comparatively simple structure of this orogene. Immediately after the Second World War, however, detailed studies started within the framework of the investigation of the Pyrenees conducted by Leiden University under the guidance of L. U. De Sitter. A great deal of interest was also shown by French geologists.

We shall give a brief summary of the most important theories and suggestions that have been put forward to account for this mountain chain. For the theories that were current before 1888, we would refer the reader to the thesis of Caralp (1888).

a. *1905: The "nappe" theory.*

According to this theory, which was propounded for the Alps and applied to the Pyrenees by Leon Bertrand, the area north of the axial zone consists of four piled-up "nappes" of which the upper two consist of Palaeozoic and crystalline formations; the satellite massives in front of the axial zone represent the lower "nappes". Bertrand is of the opinion that the calcaire métallifère is Devonian in age and that its position is accounted for by complicated overthrusting.

b. *1930: The "chaîne de fond" theory.*

Taken from the theory of Emile Argand by Jacob (1930) and Casteras (1934). A "chaîne de fond" is a mountain chain consisting of a core strongly folded at an early stage hardened by granite intrusions and migmatisation and covered with younger sediments. The core is rigid in contrast to the mobile cover. Later orogeneses, being unable to fold this hardened core, caused longitudinal faults dividing it into blocks.

According to Argand, the folded, hardened and finally levelled core may have been structurally active at various times but has had only a slight influence on the new structural style. Folding of the cover independent of the core is possible ("plis de couverture"), but in most cases the core and cover were folded together ("plis de revêtement").

c. *1934:* H. Boissevain in his thesis emphasises the plastic character of the Silurian during the folding, and compares it with the plastic behaviour of the Keuper. In an article on the Silurian of the "Vallée de la Pique" Destombes and Vaysse (1947) came to the same conclusion as Boissevain.

d. *1951:* Fourmarier assumes a cleavage front. The stratigraphical level reached by this front differs strongly owing to variations in static pressure.

The cleavage is mainly symmetrical in relation to the mountain chain and more or less parallel with the axial plane of the fold. In the Pyrenees this symmetry is expressed in a systematic southern dip of the cleavage in the northern flank and a northern dip in the southern flank. The cleavage originated during an early Alpine (Austrian?) folding phase.

e. *1956:* De Sitter (1956-a). The Pyrenees originated from a Devonian geosyncline which underwent strong and complicated Hercynian folding. The striking symmetry of this mountain chain is largely due to its intercontinental position between two stable blocks — the Aquitanian block in the north and the Ebro block in the south.

Later stages of the Hercynian orogene, which caused mainly faulting, seem to have had their greatest effect in the peripheries: a centrifugal movement continuing with the formation of Lower Cretaceous basins. An early Alpine folding phase and subsequent filling of Upper Cretaceous marginal troughs, is followed by the Pyrenean folding phase. The Alpine orogenesis appears as a direct continuation of the Hercynian folding; they seem closely interdependent.

De Sitter (1956-a) wrote on p. 232—233: "It seems reasonable to see the symmetry, the close relationship between the Hercynian and Alpine orogeneses, and the centripetal-centrifugal growth as three different characteristics of one fundamental property of this mountain chain: its intercontinental position. As both its flanking blocks, the Aquitanian block and the Ebro block, have roughly equivalent functions, it seems logical that the intervening orogene is symmetrical. It starts to develop as a central basin and, once this has been consolidated by folding, continues its growth by encroaching north and south on these flanking blocks".

B. INTRODUCTION

The Pyrenees are structurally divided into a Hercynian-folded central part, the "axial zone", flanked by Mesozoic "internal zones", which in turn are flanked by "marginal troughs" (Fig. 16).

The North-Pyrenean internal zone is separated from the axial zone by a fault-system known as the North-Pyrenean fault. This internal zone contains Palaeozoic nuclei, or "satellite massives".

The South-Pyrenean internal zone, also known as the Nogueras zone, is much narrower than its northern counterpart. Both internal zones were intensely folded during the Alpine orogene superimposed on the Hercynian folding. The marginal troughs are filled with Upper Cretaceous and Tertiary deposits about 5,000 m thick in the north and 3,000 to 4,000 m in the south.

The configuration of the southern basins extending southwards from the Hercynian mountain axis, i. e. the centrifugal movement indicated by de Sitter, can be deduced from the successive axes of thickest sedimentation from Lower-Cretaceous to Oligocene, which is more than 3,500 m thick further south.

The entire region under discussion is included in the axial zone, which is untouched by any Alpine distortion except for faulting or reactivation of Hercynian faults. This can be proved for the Baños fault, which was active after the deposition of the Trias.

The folding of the Devonian north and south of the axial zone is more concentric in character, as shown by the large and relatively simple rounded

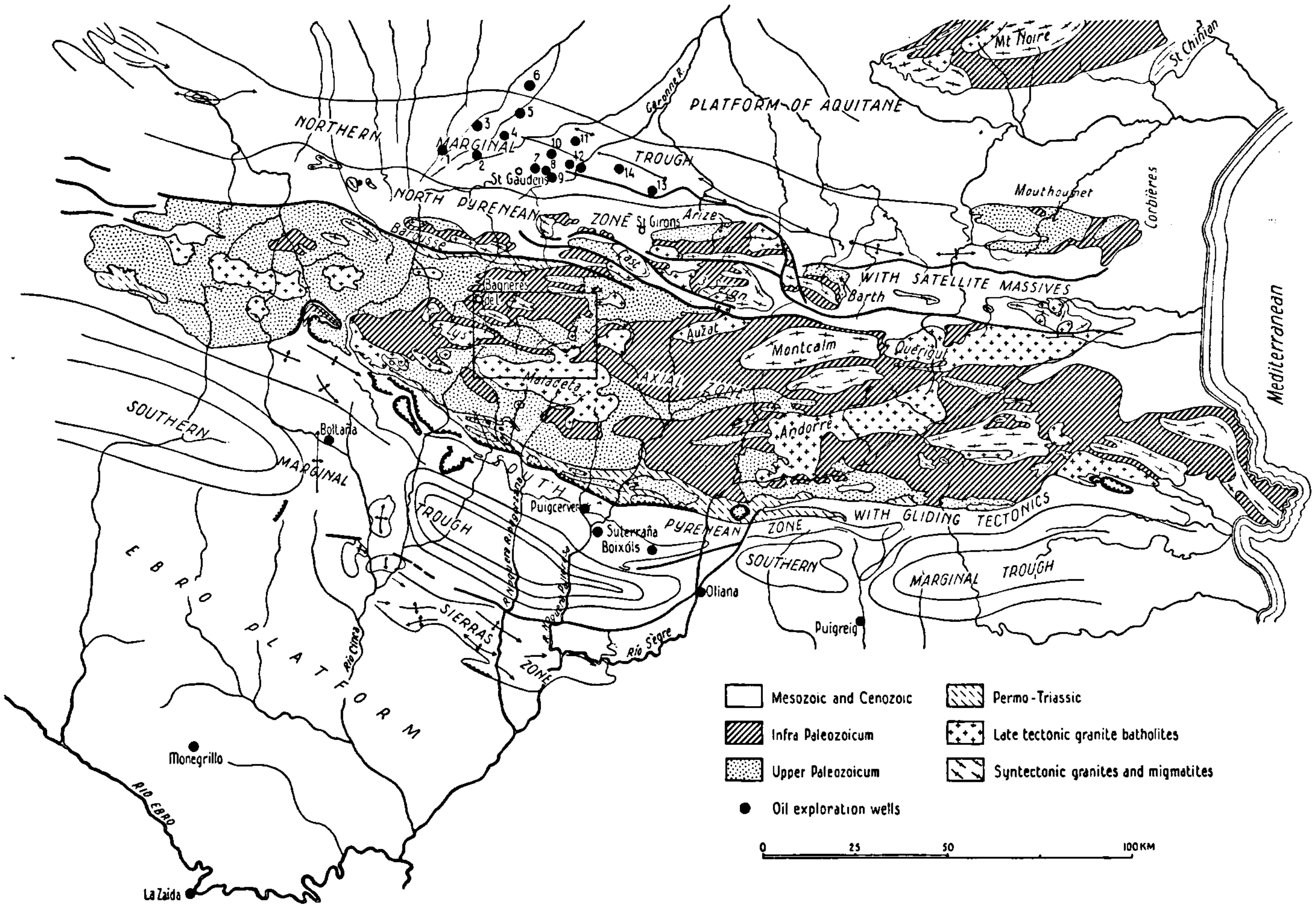


Fig. 16. Structural sketch map of the Central and Eastern Pyrenees. (mainly after de Sitter, 1956)

Data on thickness of Mesozoic and Cenozoic sediments are obtained from records of oil exploration wells published in the A. A. P. G. 1950—1958.

1. Clarens
2. St Planchard
3. Gensat
4. Lespuque
5. Mondilban
6. Puymaurin
7. Liéoux
8. Castillon
9. Laberte
10. Roupiary
11. St Marcet
12. St Martory
13. Richou
14. Plagne

folds. At the borders of the axial zone the folds become steeper and more compressed, generally showing cleavage. They are still quite large and often accompanied by secondary folds.

In the central part of the axial zone, where our area is situated, the folding is of the pure cleavage type. Here the folds are smaller and isoclinal, generally accompanied by secondary folds.

The strong folding of the Devonian is accentuated by its detachment along the Silurian slates from the Cambro-Ordovician basement.

The type of folding in the area described is closely associated with:

1. its position in the centre of the axial zone, i.c. in the centre of the Hercynian orogene.
2. the lithological sequence of the Palaeozoic sediments.

As the Silurian forms an adequate boundary between two types of folding, that of the Cambro-Ordovician basement and that of the Devonian, we shall treat these large stratigraphical units separately in our tectonical description.

The cores of the largest anticlines consist of Cambro-Ordovician. Thus we distinguish a Northern Anticline in the frontier region, associated in the south with a highly metamorphic dome-like structure, the Bosost Dome, which is separated from the Northern Anticline by a large dislocation called the Bosost fault. Faults and cracks, mostly longitudinal, are prominent in the centre of the Northern Anticline.

North of the Maladeta s.l. the southern flank of the Central Anticline is cut off by the Jueu faults between the Rio Jueu and Rio Negro. The Negro fault separaes the north flank from the Devonian. Despite the cover of Devonian limestones and Silurian debris, the presence of Cambro-Ordovician can be deduced from the large size of an anticline the north flank of which is exposed south of the Maladeta s.s. This anticline will therefore be described together with the Northern and Central Anticlines.

The Marimaña dome appears in the eastern part of our map; major faults, the Baguera, Pudo and Baños faults, affect its south flank.

The Devonian is mainly composed of limestones (Basal limestone), slates (Entecada slates) and sandstones (Las Bordas sandstones), each with their own folding characteristics. Large faults are not known in the Devonian, except the Juan Martin fault north of the Salardu granodiorite.

A faulted and downwarped Carboniferous syncline is separated from the Central Anticline by the Jueu faults, and from the Maladeta s.l. by the Maladeta fault.

Clearly-exposed structures are rare in this strongly-folded area, but an attempt has been made to illustrate the cross-sections with fold characteristics sketched from photographs. The location of the sections is given in Fig. 47.

C. CAMBRO-ORDOVICIAN

a. *The Northern Anticline*

Owing to the well-differentiated sediments of the Northern Anticline its structures are well-known, although folds can rarely be seen in a single outcrop. There is no doubt, however, that the Northern Anticline consists of successive structures like those of the May de Bulard anticline (Fig. 17), which illustrates the simplicity and size characteristic of the Cambro-Ordovician structures.

The southern and western continuations of the May de Bulard anticline are shown in the cross-sections of Fig. 18. Further west, where the calcaire métallifère is absent and there has been metamorphic influence, the structures are less clear. The plunge is generally eastwards, greatly increasing in the May de Bulard area.

Faulting is prominent; the most important is the Bosost fault which can easily be recognized in the field. The Liat fault is indicated by thick quartz veins and remnants of Silurian slates preserved in the fault scarp. On aerial photographs the fault shows itself as a pronounced scar in the landscape extending for quite a distance. Quartz veins and highly disturbed sediments (generally showing a higher degree of metamorphism) are the usual fault indications.

The faults are mostly longitudinal and more or less parallel to the main

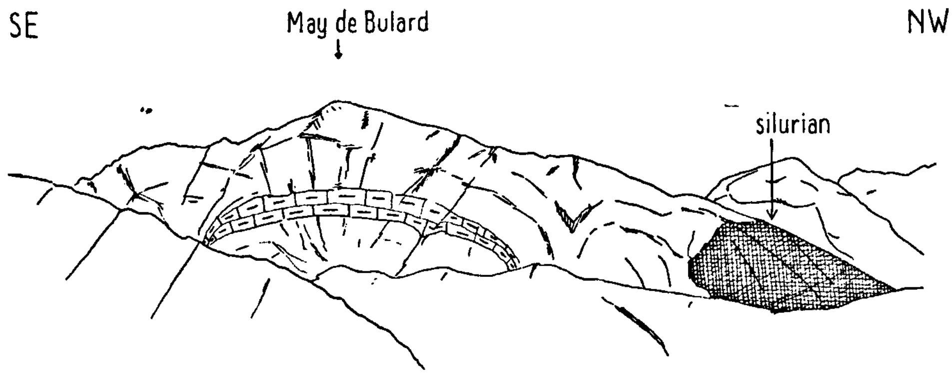


Fig. 17. The Cambro-Ordovician Anticline of the May de Bulard. After photograph.

structural trend. There are also transverse faults such as the faults of the Iñola and Orle extending from NNW to SSE, and the faults north of the Plâ de Tur which run from NE to SW.

Cleavage, although nearly always difficult to distinguish in the field, shows clearly in thin sections. Metamorphism is expressed by the occurrence of chiastolite in the Silurian slates, which especially in the west attain large sizes (up to 10 mm). This is more likely to be associated with the highly metamorphic areas of Bosost and the Rio Toran rather than with the Riberot granite west of the Orle river (not shown on the map) as was suggested by Destombes (1958).

Generally the sediments are highly silicified. The silification also accounts for the great number of quartz veins, which are often associated with faults and cracks. Mineralization, also bound to faulting, seems to be associated with this process. A distinction can be made between hydrothermal and pneumatolitic-hydrothermal mineralization.

The calcaire métallifère acted as a resorbing rock, whereas the carbonaceous Silurian slates occasionally had a screening effect. Hydrothermal minerals have been absorbed through metasomatism by the limestone. Ore

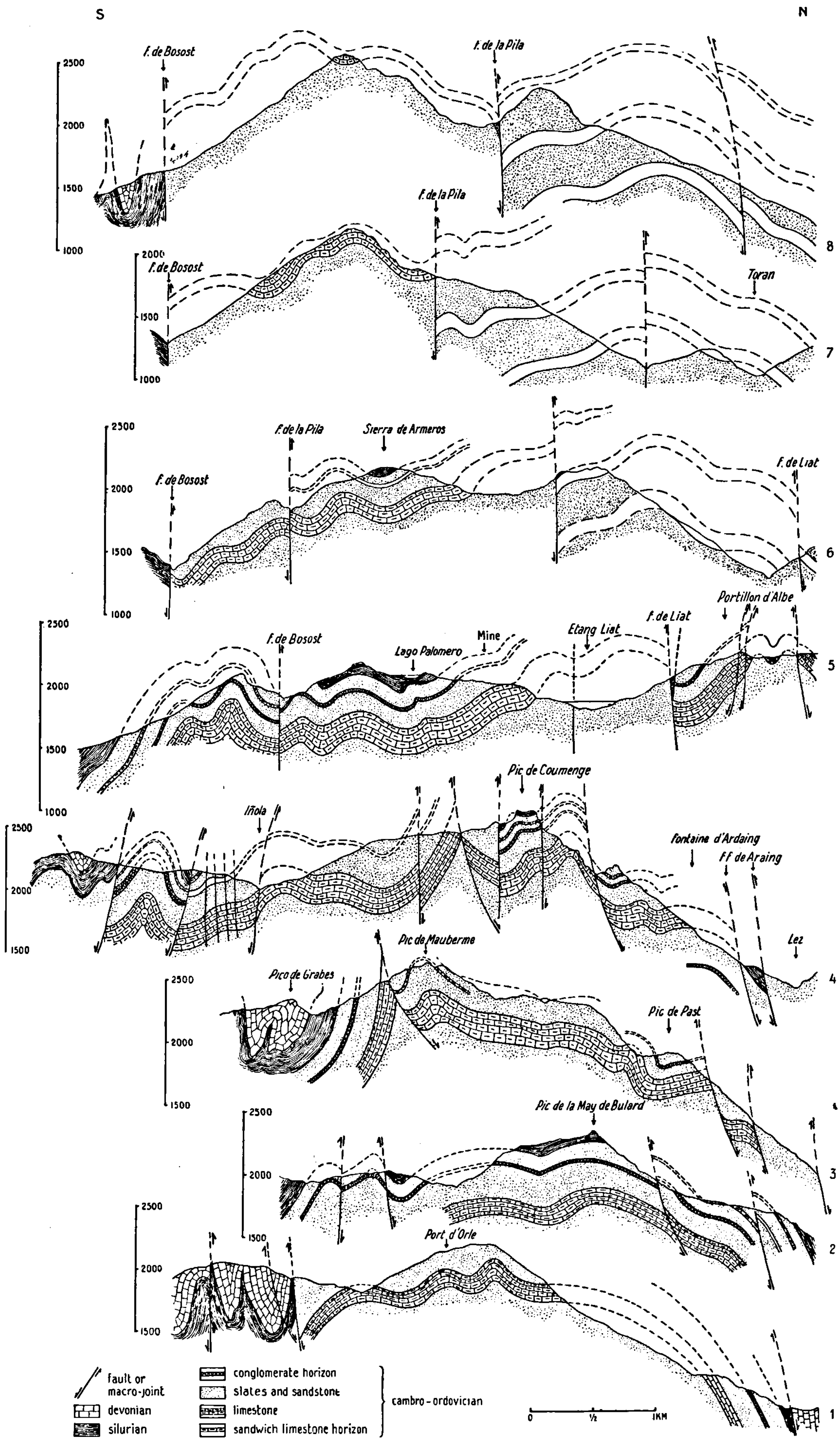


Fig. 18. Cross-sections through the Northern Anticline

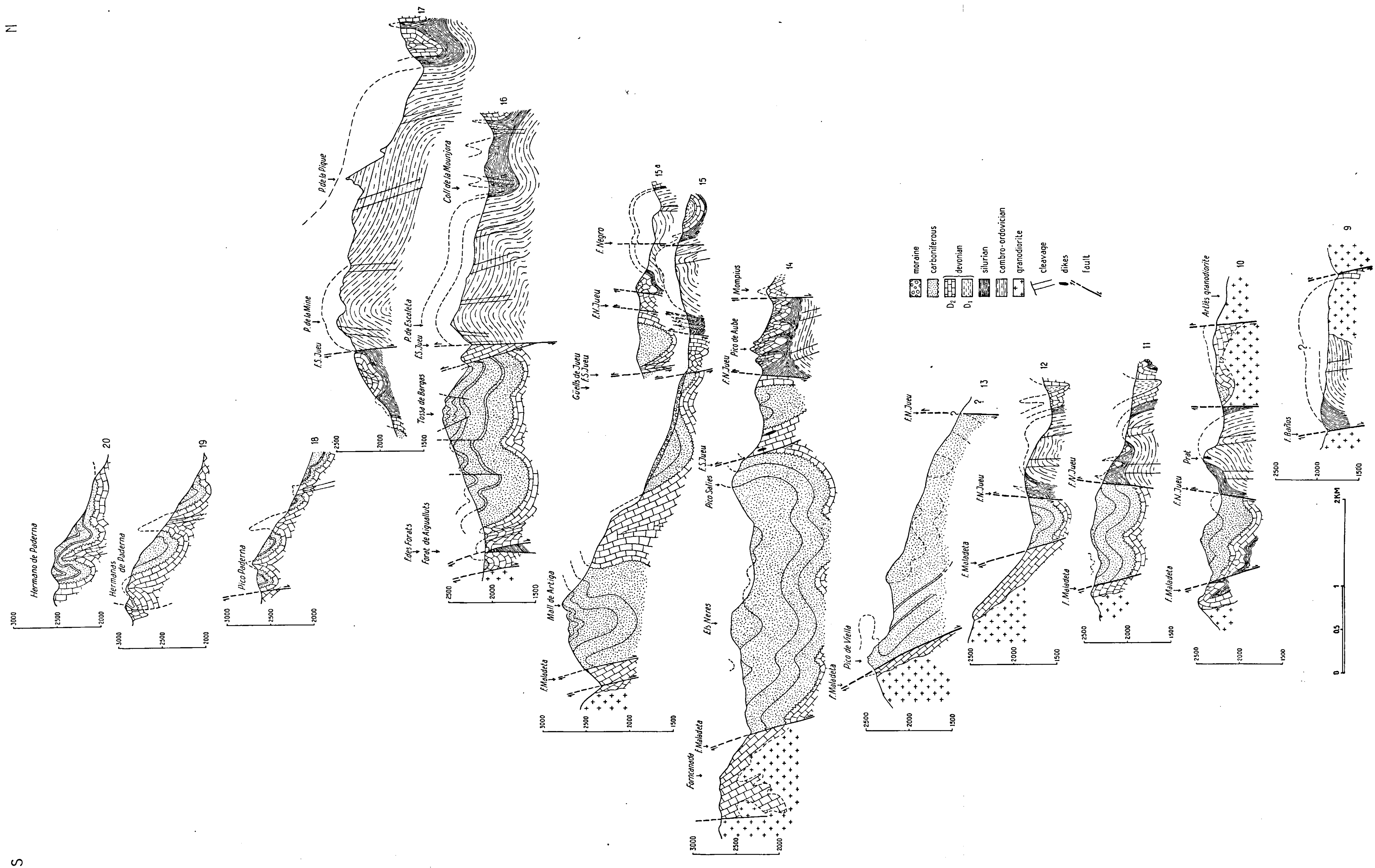


Fig. 19. Cross-sections through the Central Anticline and the Carboniferous

bodies mainly consisting of quartz with galenite and sfalerite were formed, occasionally accompanied by pyrite bodies. Of this type are the ores of the abandoned mines of Victoria (468/52), Margarita (469/53), May de Bulard (487, 488/58) and Bentaillou, which is approximately 3 km north-east of Liat, outside the map.

Along faults and cracks occur many small-sized veins of pneumatolitic-hydrothermal origin, consisting mainly of quartz in combination with pyrite, arsenopyrite, chalcopyrite, pyrhotite.

This process of mineralization continued after the development of faults, cracks and the knicking of cleavage. Very likely the mineralization associated with faults and cracks is of Hercynian origin; galenite and sfalerite ores in the Pyrenees have not penetrated into Mesozoic rocks.

b. The Central Anticline

North of the Maladeta s.l. the youngest sediments have been preserved in a tectonical depression, flanked to the north by a steep Cambro-Ordovician anticline, the "Central Anticline". Cross-sections through both structural units are given in Fig. 19.

In the drainage area of the river Lys (457/50) the Central Anticline consists of two parallel anticlines, narrowing towards the river Jueu and there limited to the north and to the south by faults. East of the river Jueu an excentric culmination of Cambro-Ordovician crops out north of the Negro fault; evidence of an eastwards continuation of this fault has not been found.

The eastward plunge of the structures is much more marked in the Prat area (4°.32′/42°.40′). Moraine material obscures the country rock between the rivers Valarties and Aiguamoix, but a large anticline exposed north-east of the Baños de Tredós probably represents the eastern continuation of the Central Anticline. The northern flank of this structure is occupied by the Tredós granodiorite, which caused a broad metamorphic aureole. The southern flank is separated from the Maladeta s.l. by the Baños fault. North of the Pico de Salana a broad mylonitic zone in the granodiorite indicates the Baños fault (Fig. 20-C).

Fault indications are frequent in the Triassic sediments; the wavy surfaces of small fault planes are covered with a thin layer of chlorite (Fig. 20-D). Longitudinal faulting is prominent in the east; the arrangement "en échelon" of the faults on the northern flank and some of those on the southern flank as well as the sudden steepening of the plunge west of the river Valarties might indicate transverse faulting parallel to the valleys. The faults are probably late-Hercynian, and some post-Triassic (Baños fault), reactivated during later (Alpine) epirogenetic movements.

The folding is different from that of the Northern Anticline. A section given in Fig. 57-a as seen from the east illustrates the great compression of the predominantly pelitic sediments. Shortening may be as much as 60 % or 70 %.

Cleavage is generally well-devedoped; knicked cleavage is a common feature. When approaching the Port de Venasque the sediments become more and more silicified, and obviously this silification process is later than the knicking of cleavage. Apparently the silification is associated with the southern Jueu fault, which runs south of the Port de Venasque.

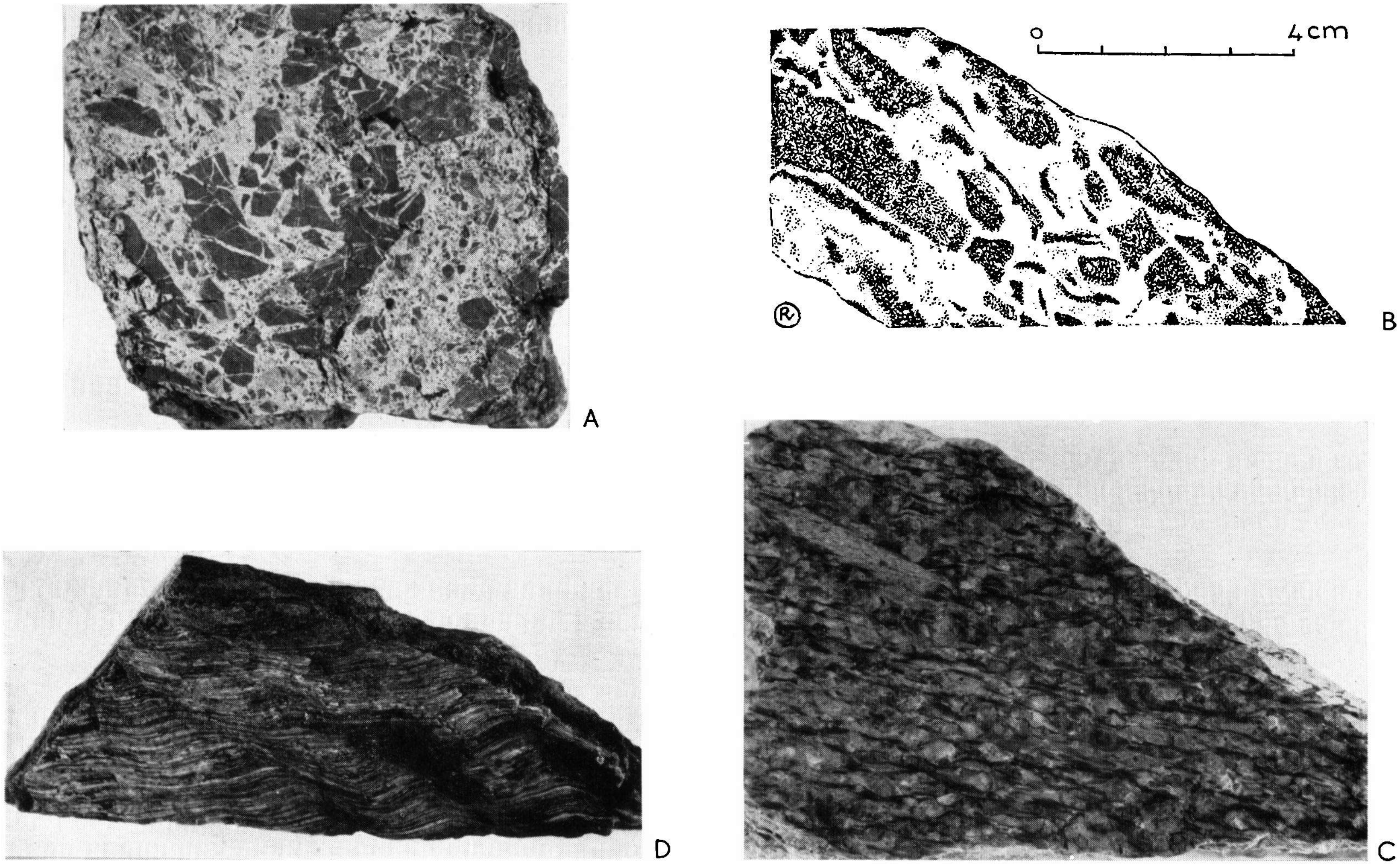


Fig. 20. Fault indications

A) Fault breccia of the Maladeta fault, south of the Collado de Toro
B) Fault breccia from small faults in the Las Bordas sandstones (after photograph)
C) Mylonite from a mylonite-zone west of the Pico de Salana
D) Fault plane in Permo-Trias sandstone covered with a 2 mm thick layer of chlorite

c. *The Southern Anticline*

A cover of Silurian debris obscures the underlying Cambro-Ordovician structure, as is drawn in the cross-sections of Fig. 21. As the upper part of the Rio Valibierna is approached the large size of the anticline is immediately noted (Fig. 22).

The highly intercalated Devonian basal limestones follow the anticlinal

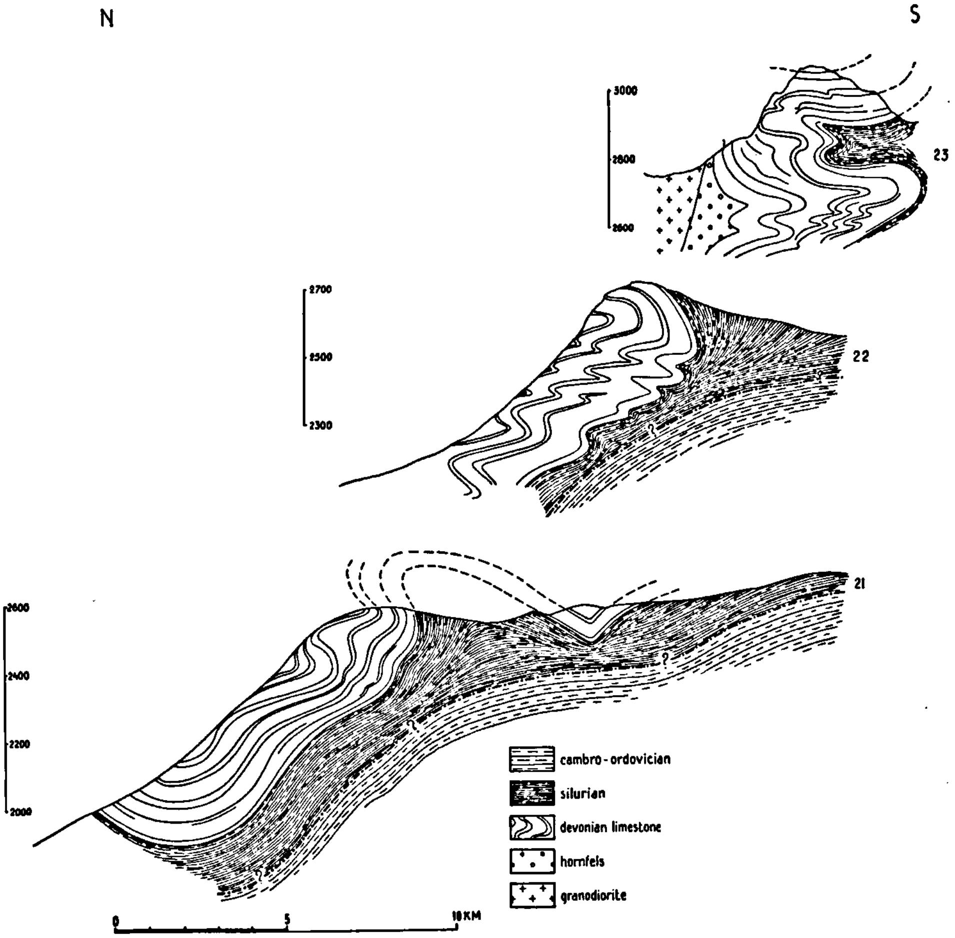


Fig. 21. Cross-sections through the north flank of the Southern Anticline

shape (Fig. 21, section 21) for a certain distance occasionally forming small recumbent folds (Fig. 23) or even a cascade of much larger folds (Fig. 24). Gently-folded isolated Devonian synclines are found on the crest of the anticline.

The piling-up of the Devonian against the granodiorite and the gliding tectonics on the northern flank suggest development later than the main Hercynian folding phase, probably during, or after, the intrusion of the granodiorite,

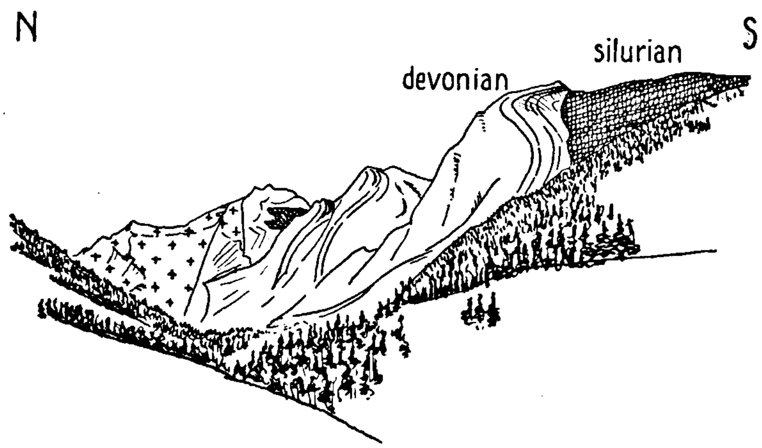


Fig. 22. North flank of the Southern Anticline exposed in the Rio Valibierna. After photograph

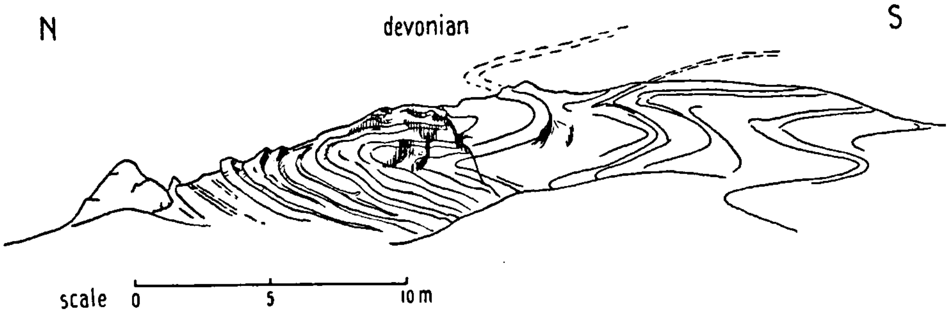


Fig. 23. Gliding structures developed in Devonian limestones on the north flank of the Southern Anticline. After photograph

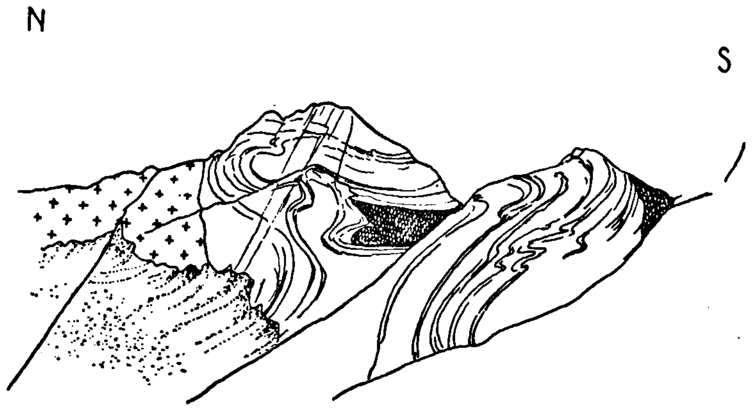


Fig. 24. North flank of Southern Anticline, detail of Fig. 22. After photograph

d. The Bosost Dome

The Bosost fault separates two metamorphic areas: the symmetrical culmination of the Vallée de Burbe-Portillon (461/53) in the south, and that of Bosost in the north (Fig. 25, section 25).

The anticlinal axis of the first area is well-exposed in the Colle de Portillon (463/53); the anticlinal axis of the other runs through the village of Lès. The Cambro-Ordovician although strongly micro-folded, forms relatively large folds, in contrast to the intensely-folded overlying Devonian (Fig. 57-b, section 24).

Lineations measured in an area extending from La Bordeta (highly metamorphic sediments) to the Rio Barados (slightly metamorphic sediments) run roughly parallel (Fig. 26 *). Zwart (1958) measured parallel lineations in

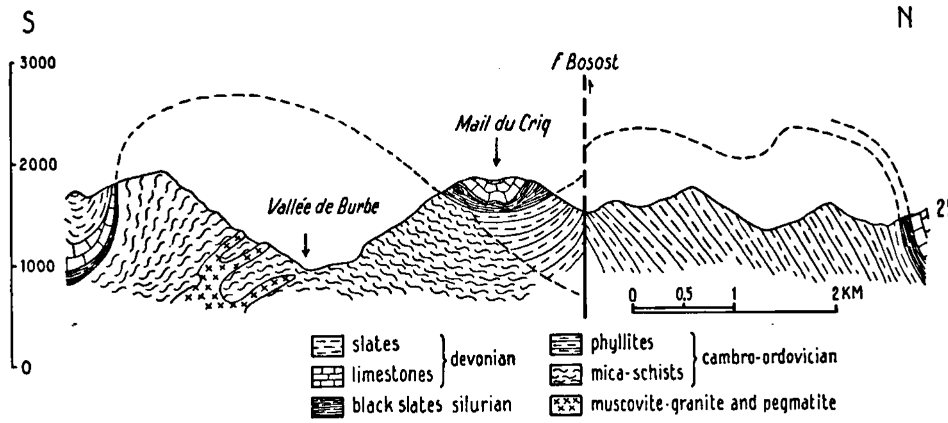


Fig. 25. Cross-section through the Bosost Dome

schists surrounding the granites north of La Bordeta and occurring within these granite bodies.

This parallel arrangement of lineations differs from the arrangement of lineations around the granodiorites, which run more or less parallel to the granite boundaries. This indicates another mechanism of emplacement of the Bosost granites different from the "shouldering aside" of the granodiorites.

e. The Marimaña Dome

Cross-sections of this region are given in Fig. 27.

The intruding Marimaña granodiorite caused doming of the already folded surrounding sediments, accentuating the western plunge of the structures (Fig. 28). Structures can hardly be detected in the monotonous black slates. One of the thin limestones, however, is folded into an anticline as can be observed further to the east, where this limestone is connected to the calcaire métallifère surrounding the granodiorite.

*) The fold axes south of Arrès which deviate from the main direction are measured in metamorphic limestones.

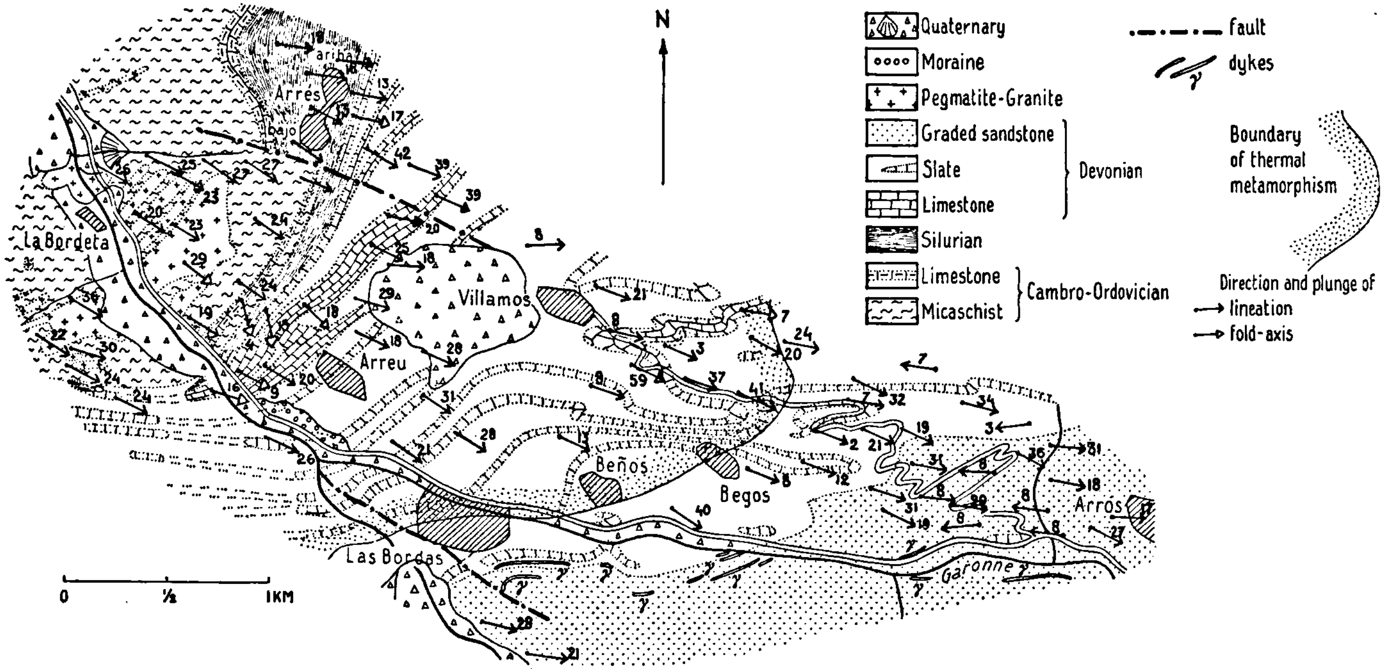


Fig. 26. Map of liniations and folded-axes measured in the transition zone of the highly metamorphic Cambro-Ordovician in the west and the slightly metamorphic Devonian in the east

It is not known whether the other thin limestones in this area are also connected to the calcaire métallifère, but we believe they represent restricted limestones of a stratigraphically higher level.

The boundary between the black slates and the calcaire métallifère is a normal stratigraphic one, as is indicated by an accompanying thin-bedded limestone horizon at a constant distance from the calcaire métallifère, which probably thins northwards.

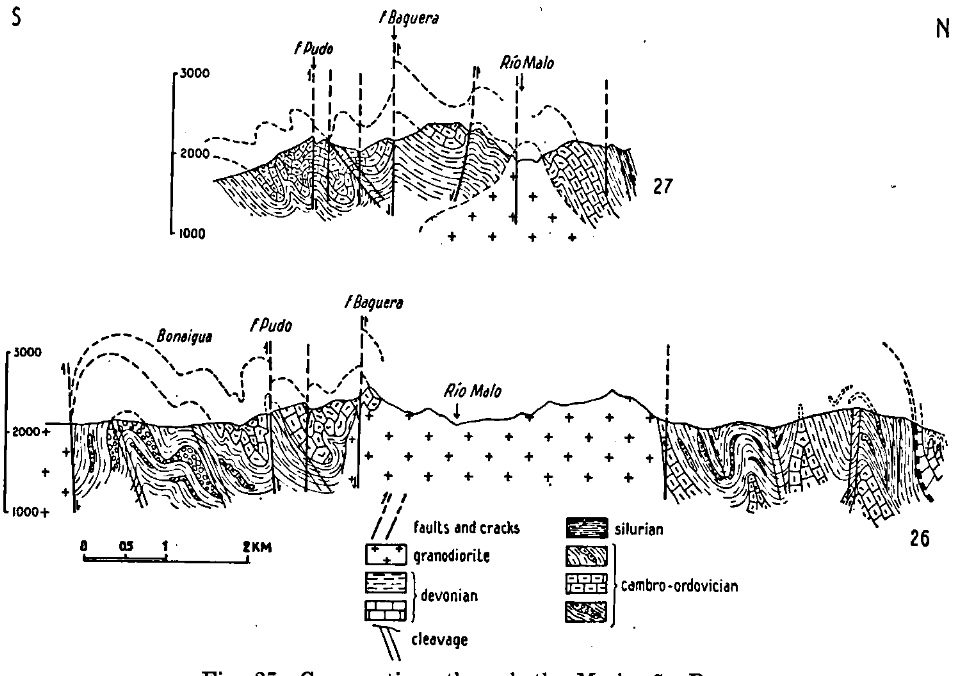


Fig. 27. Cross-sections through the Marimaña Dome

The boundary between the Cambro-Ordovician and the Devonian is formed by a steep flexure. The carbonaceous Silurian in between is very thin and may even be absent locally.

South-west of the Marimaña a wedge of black slates which penetrates deeply into the limestone is connected to metamorphic rocks which are undoubtedly older. Here a brecciated zone and thick quartz veins are indicative of the Baguera fault. Isolated limestones with southward-plunging structures are cut off by the intrusion. To the south limestones cover an extensive area in which structures can hardly be detected.

Sharp lines on the aerial photographs were not identifiable as faults. They most probably represent cracks. Many small springs situated along these cracks occur in the limestones of the Ruda valley.

The Pudo fault is indicated by linear patches of slate in the calcaire métallifère, and locally by thin quartz veins. East of the watershed, water which disappears in sink-holes along the Pudo fault, reappears in the Fuentes de Ruda in its western continuation.

The sediments of the Puerto de la Bonaigua represent the oldest formation in this area. The limestones and conglomerates probably form restricted horizons situated below the calcaire métallifère as is drawn in cross-section 26 (Fig. 27). They are separated from the Maladeta s.l. by the eastern continuation of the Baños fault. Along this fault carbonaceous slates which have been attributed to the Silurian occur in one place. This indicates a large anticlinal structure of the Bonaigua series confirmed by the occurrence

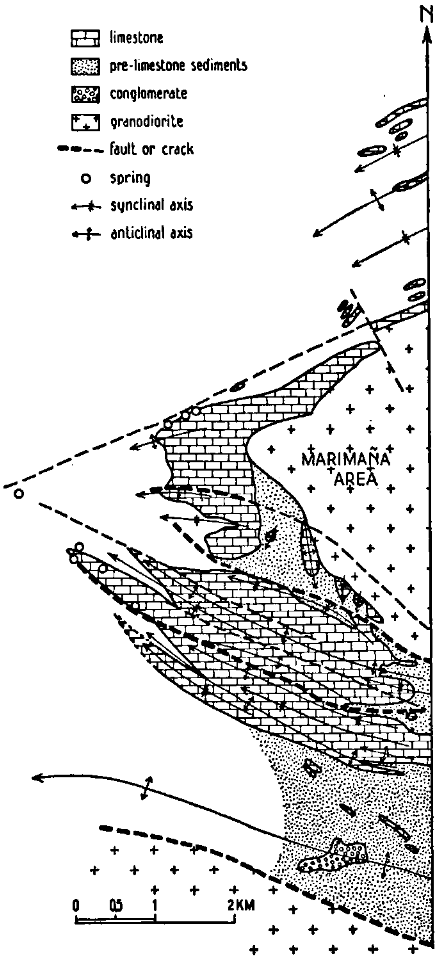


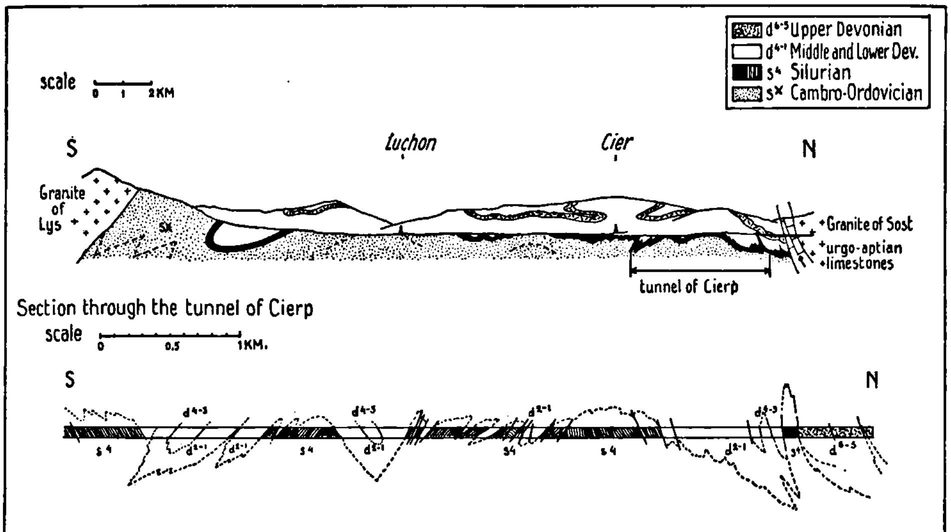
Fig. 28. Structural sketch map of the Marimaña Dome

of a Cambro-Ordovician anticline west of the Rio Garona de Ruda. Probably the Bonaigua anticline represents the eastern continuation of the Central Anticline.

D. SILURIAN

Structurally the Silurian, as a particularly incompetent horizon, can be described as the lubricating horizon which caused detachment of the Devonian cover from the Cambro-Ordovician basement with all its accompanying complications.

A good example is given by Destombes and Vaysse (1947, Fig. 29) from a tunnel section along the Pique north of Bagnères de Luchon. They assume that during the Pyrenean folding phase the soft non-metamorphic Silurian played a plastic role, comparable with that of the Keuper in relation to the Hercynian core and the Mesozoic cover. Thus it is possible for wedges of Silurian slates to break through the overlying layers locally (Fig. 30) sometimes dragging parts of these layers along. Such wedges alternating with marmorized Devonian limestone were found during the building of a tunnel, 1 km north of the Guëlls de Jueu.



Simplified section through the Pique Valley with a detailed tunnel section. After J.P. Destombes and A. Vaysse, 1947

Fig. 29

According to Destombes and Vaysse Silurian slates are often recumbently folded on top of, or against, the Cambro-Ordovician basement, becoming sub-vertical between the Devonian limestones.

This mobility of the Silurian accounts for the completely disharmonic folding of the Devonian as compared with the Cambro-Ordovician. Great variations in thickness and abundant signs of strong deformation like slickensiding have often been observed during field work. The mobility of the Silurian is probably due to the very fine granularity and high carbon content of its sediments, resulting in a considerable difference in competency as compared with the Cambro-Ordovician and Devonian.

The steep folds of the Devonian basal limestone are occasionally exposed (Fig. 31) and the large Cambro-Ordovician structures are well known from the Northern Anticline (Fig. 18).

South of the Cabeza de Coll thin Devonian basal limestones wedging out towards the east and west alternate with thin Silurian and Devonian slates, thus forming imbricated structures. Along thrust faults Silurian slates have penetrated deeply into the overlying Devonian limestones and slates. Remnants of Silurian slates preserved in fault planes are sometimes common, causing difficulties in the lithostratigraphic interpretation.

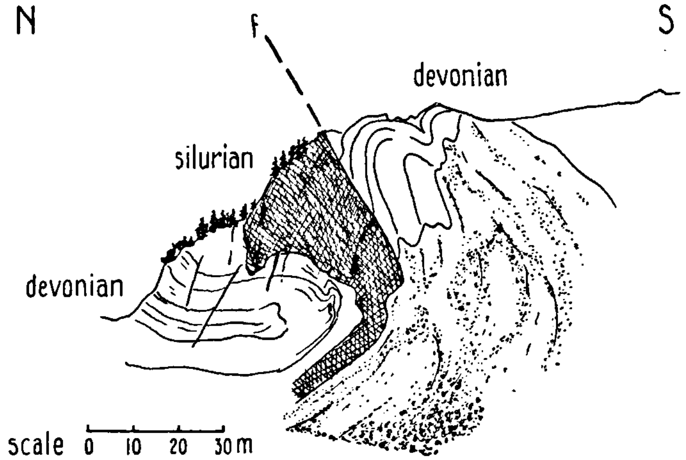


Fig. 30. A wedge of Silurian black slates in faulted contact with the overlying Devonian limestones, exposed north of the Sierra de Betlán. After photograph

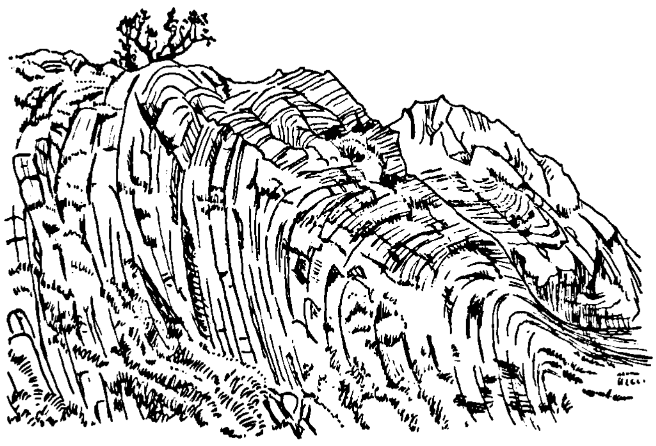


Fig. 31. Folds of relatively large size in Devonian limestone, Pic de Secoube area. After photograph

E. DEVONIAN

a. Limestones

The Basal limestone shows a great variety in type of deformation; micro-folding is frequent (Fig. 10). Micro-folds in intercalated limestones clearly depend on the thickness of the alternating layers. The greater the distance

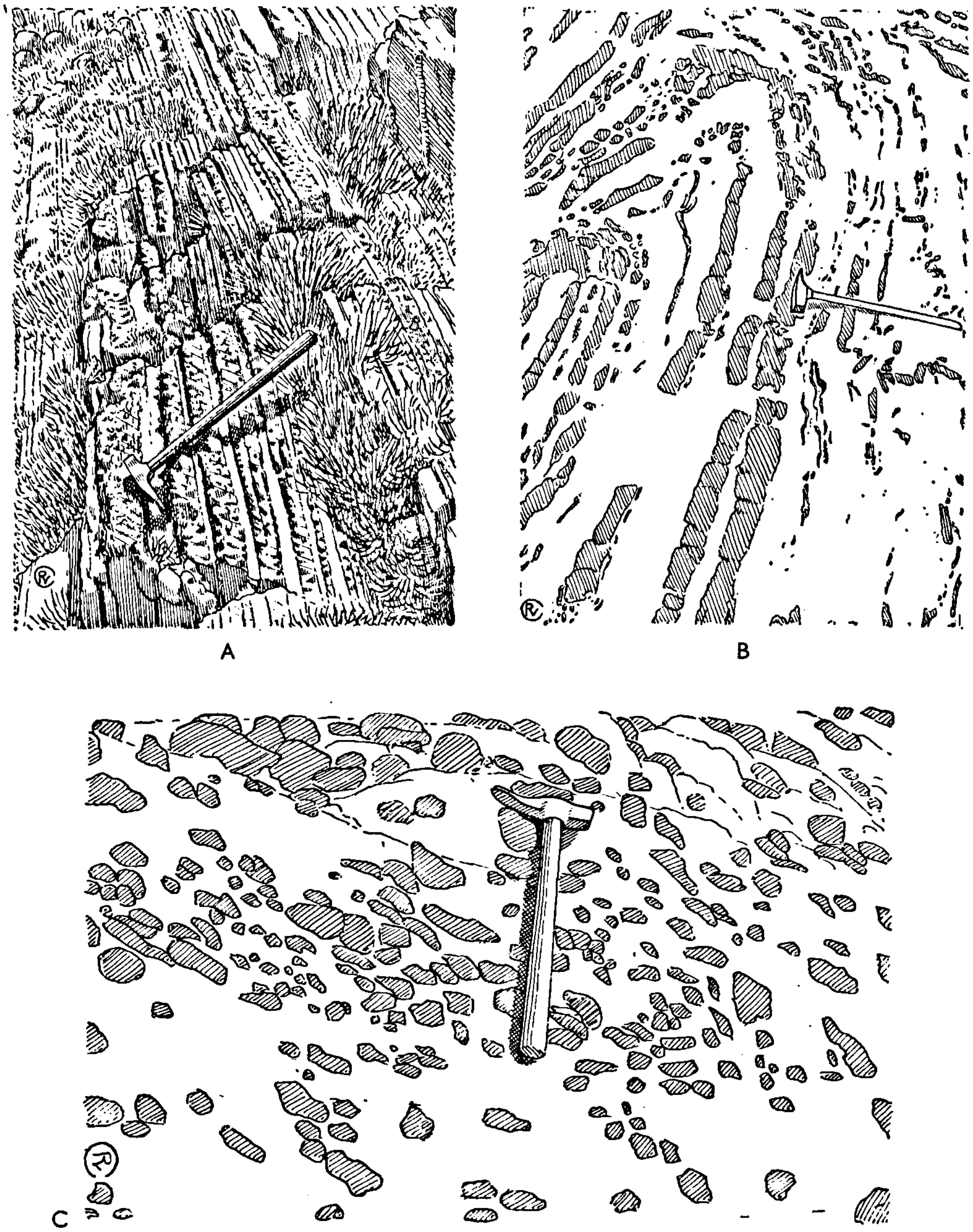


Fig. 32. Three successive phases of increasing flow in limestone as seen in a limestone-slate alternation:

A) Undisturbed; cleavage is well developed but already vague in the limestone

B) Flow structures. Cleavage is no more visible

C) Chaotic arrangement of slate fragments in limestone. No sign of former cleavage or flow structure is left.

After photograph

between the intercalations (and the thicker the intercalations), the greater the micro-folds, ranging from 5 cm to several metres.

Micro-folds often show characteristics of flow, particularly in marbles. Minute shale fragments arranged along the cleavage planes and later micro-folded are the only indications of this secondary kind of folding.

It has been observed that, towards the centre of a large Devonian syncline, a limestone-slate alternation showing a well-developed cleavage looses this cleavage in the limestones. Higher up in the sequence the slates are fractured into small particles, owing to flow of the limestone, but structures are still visible. Finally a completely chaotic arrangement of slate fragments in the limestones is observed (Fig. 32).

The method of determining the shortening with the help of broken shale fragments in limestones is not very reliable, and varies within short distances.

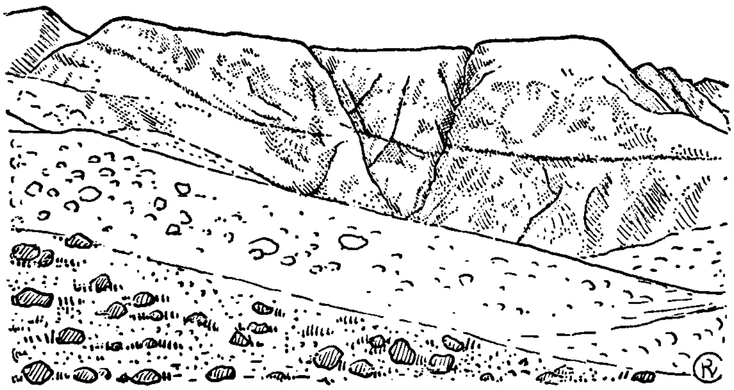


Fig. 33. Normal faults of the Pico de Parros. After photograph

Not all intercalated limestones show such intensive micro-folding; in places the limestones in the Sierra Negra are arranged in folds of much greater width.

More or less pure limestones have larger fold characteristics (Fig. 31) such as are also shown in an alternation of relatively thick limestones, sandstones and slates (Fig. 57-c).

b. Slates

Folds in slates can only be observed when they contain intercalations of more competent layers, such as are usually observed in the Las Bordas sandstones. In the Entecada slates bedding is completely obliterated by cleavage. Folds in such thick uniform slate formations should be relatively large owing to their low resistance.

Faults and cracks are frequent in the Entecada slates, but they are mostly parallel to the cleavage and therefore difficult to detect. Such faults and cracks in slates have been studied in a tunnel section east of the river Jueu; some of them were partly cemented with calcite. Only near the Pico de Parros (486/42°.46′) are small-sized normal faults oblique to the main structural trend well-exposed (Fig. 33).

c. Graded Sandstones

Graded beds have their own fold characteristics: small folds of little amplitude, and frequent exceptionally steep plunges (Fig. 34). This characteristic is especially well-developed between the village of Mont and the river Iñola (Fig. 35). Folds in sandstones of somewhat larger dimensions

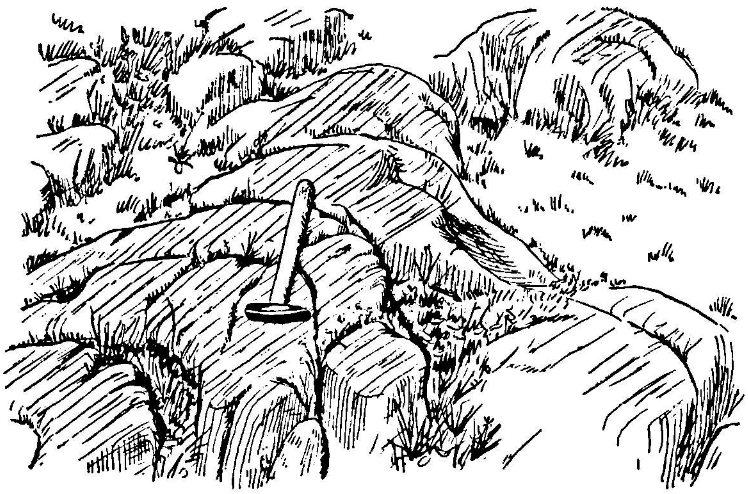


Fig. 34. Steeply plunging fold in the Las Bordas sandstones, west of the Areño. After photograph

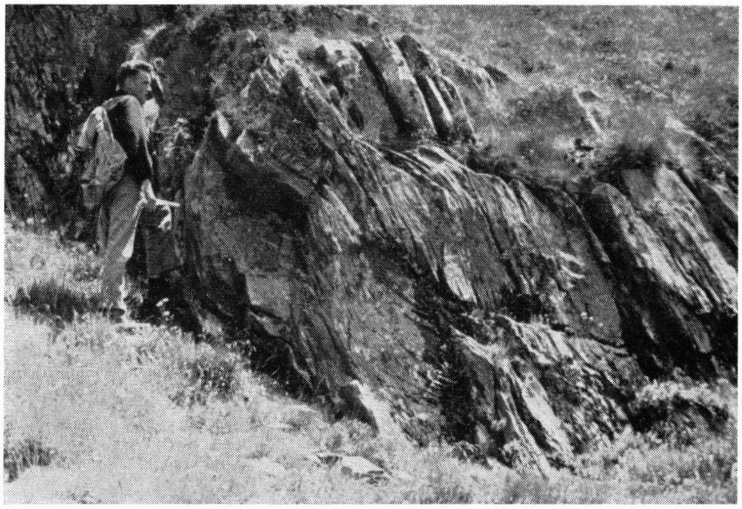


Fig. 35. Cleavage fold in graded beds. The inward bending of the cleavage in the graded beds is clearly exposed

have been observed north and west of Viella and north of Aubert (473/48) (Fig. 57-d). The shortening here may vary between 50 % and 70 %.

Large faults have not been observed but small-sized longitudinal faults occur frequently. West of Garos one of these faults was studied in detail; horizontal as well as vertical displacement was ascertained.

In areas where sandstones and quartzites predominate fault breccias are occasionally found (Fig. 20-b).

Small vertical faults perpendicular to the fold axes have frequently been noted (Fig. 36). Their origin is not clear; they were probably formed during a late Hercynian phase.

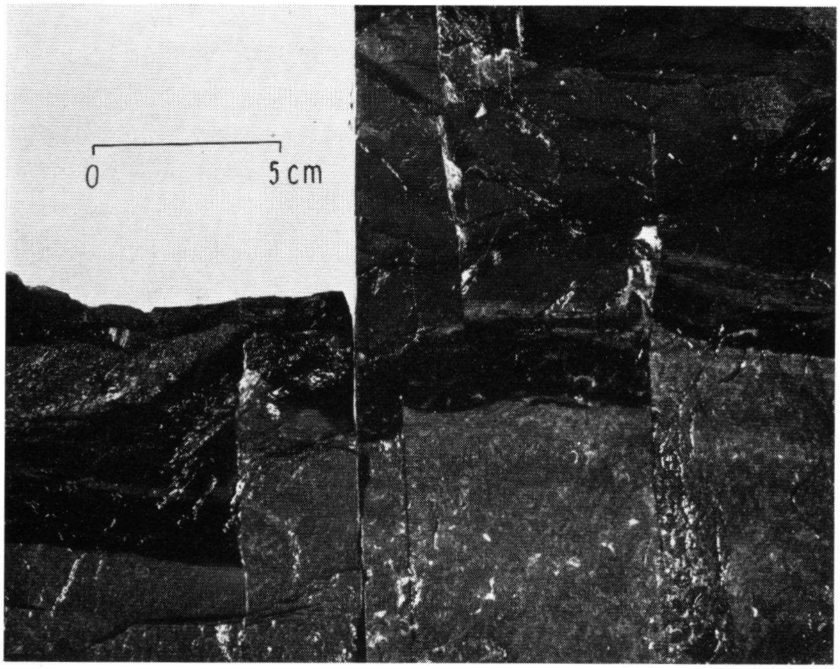


Fig. 36. Small vertical faults, perpendicular to the main structural trend developed in graded beds, north of Montcarbau

d. The Corbison and Monte Calvo areas

In this region bordered to the north by the Juan Martin fault and extending from Salardu to the eastern slope of the Corbison, a considerable differentiation in sediments can be observed. The Basal limestone is not exposed, and consequently the thickness of the Entecada slates is unknown. But the very reduced slate series exposed east of Salardu indicates a considerable eastwards thinning of those slates. The succeeding Entecada limestone and Las Bordas orthoquartzites are also reduced here, although to a much lesser degree. Furthermore, the series consists of alternating Las Bordas sandstones and slates, a thin limestone (not always present) and green Viella slates.

The kind of folding in this region can be observed north of Arties and Salardu in the deeply incised valleys running from north to south (Fig. 57-c).

The shortening could easily be measured at two successive folds. Thin quartzitic layers present in these folds do not show any appreciable thickening

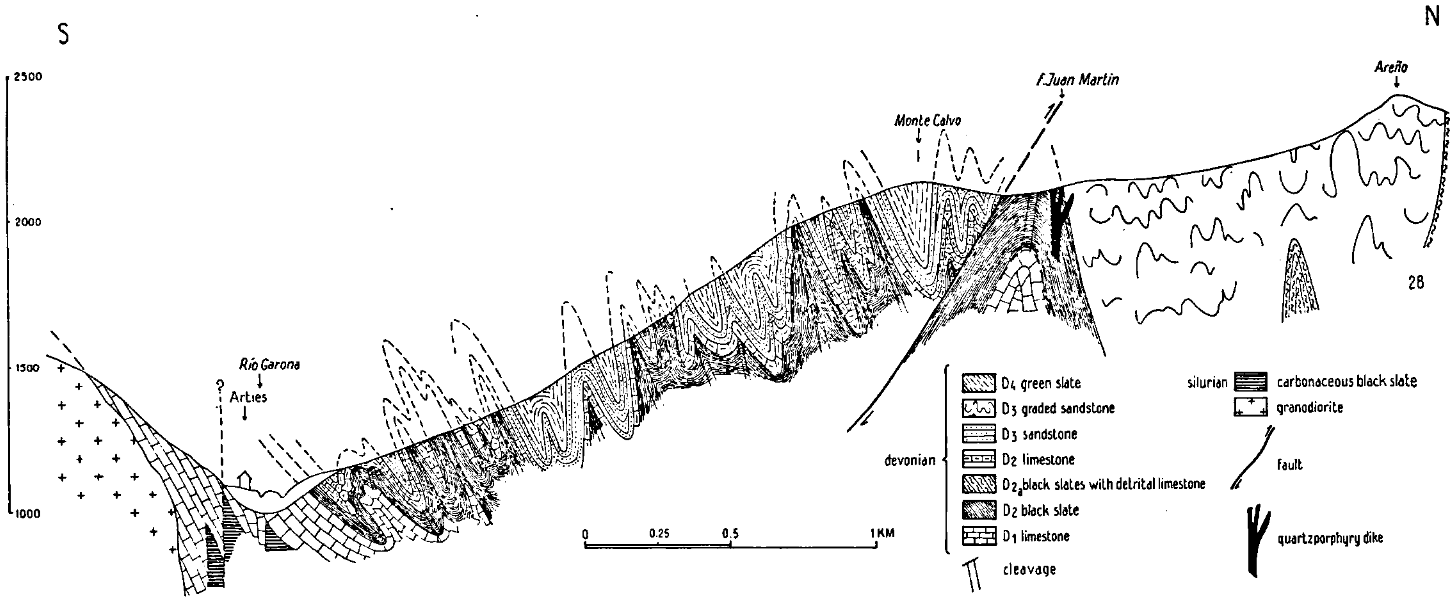


Fig. 37. Section north of Arties

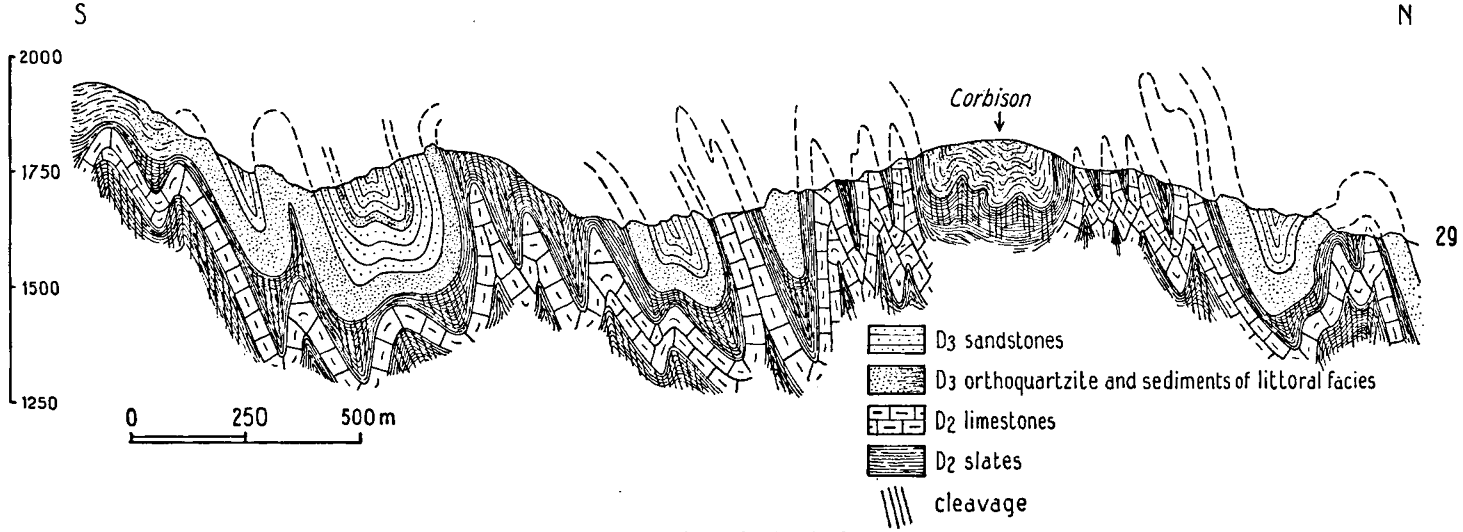


Fig. 38. Section through the Corbison area

in the anticlinal hinge. Such secondary thickening is of no importance in calculating the shortening compared with the large fold amplitude.

The following percentages of shortening were calculated: 60 %, > 60 % and 68 %. On the eastern slope of the Corbison the successive anticlinal and synclinal hinges were in most cases successfully located except in some limestones and the littoral deposits.

The Entecada limestone is not developed west of the ridge descending to Las Bordas. The pattern of the eastward-plunging structures on the western slope of the Corbison could therefore not be observed. The limestones on the east slope of the Corbison always expose the anticlinal hinges. The plunge of the structures is almost the same as the topographic slope, which gives the limestones the appearance of almost parallel beds.

In the cross-sections of Fig. 37 the folds of the Corbison area are drawn with slightly larger amplitudes than those of the Monte Calvo area, the alternating layers in the latter being thinner (Fig. 38).

Large faults, except the Juan Martin fault, are unknown. A small mylonitic zone north of the Salardu granodiorite and small isolated patches of limestone sometimes accompanied by quartz veins mark this fault. Its western extension is difficult to follow; the fault probably diminishes near the slope of the Sierra de Vilach, where indications of a less important fault are found. The continuation of the Juan Martin fault east of the Rio Iñola is indicated by the absence of the Silurian and the Devonian basal limestone along the Cambro-Ordovician slates. Further east the fault disappears under alluvial deposits of the Plâ de Bérèt. The origin of this upthrust is probably associated with the post-Hercynian intrusion of the Salardu granodiorite.

F. CARBONIFEROUS

A thick Carboniferous sequence has been preserved in a syncline in a structural depression north of the Maladeta s.l. in between the Jueu faults to the north and the Maladeta fault to the south, as is shown in the cross-sections of Fig. 19. These predominanly coarse greywackes are intensely folded, so much so that the underlying Devonian basal limestone is deeply pressed into the core of the Carboniferous syncline (sections 15, 18, 19, 20).

The relatively large folds generally plunge eastwards. Fold characteristics are given in Fig. 39.

Cleavage is best developed towards the top of the sequence, where in the centre of the syncline the folding is more intense (Fig. 40).

Shortening varies between 16 % and 27 % from bottom to top in the Mall de Artiga area, reaching a maximum of 40 % in the Tossa de Bargas (4°.22′/42°.41′).

A recumbent fold (Fig. 41-A & B) observed in the Pico de Viella (4°.27′/42°.39′) (Fig. 19, section 13) contrasts sharply with the structural style of this part of the axial zone. Here the Carboniferous is more disturbed than elsewhere. This is indicated by the sharply increasing plunge of the structure (Fig. 41-C) and by the occurrence of granite dykes. It is unlikely that this fold results from the main Hercynian folding phase, but it may have been caused by the "shouldering aside" of the intruding granodiorite.

The flanking faults are clearly defined in the field. Fault characteristics are given in Fig. 20.

Small sized faults resulting from the main Hercynian folding phase occur frequently. They produce the same fault breccias as those observed in Devonian sandstones and quartzites (Fig. 20-b).

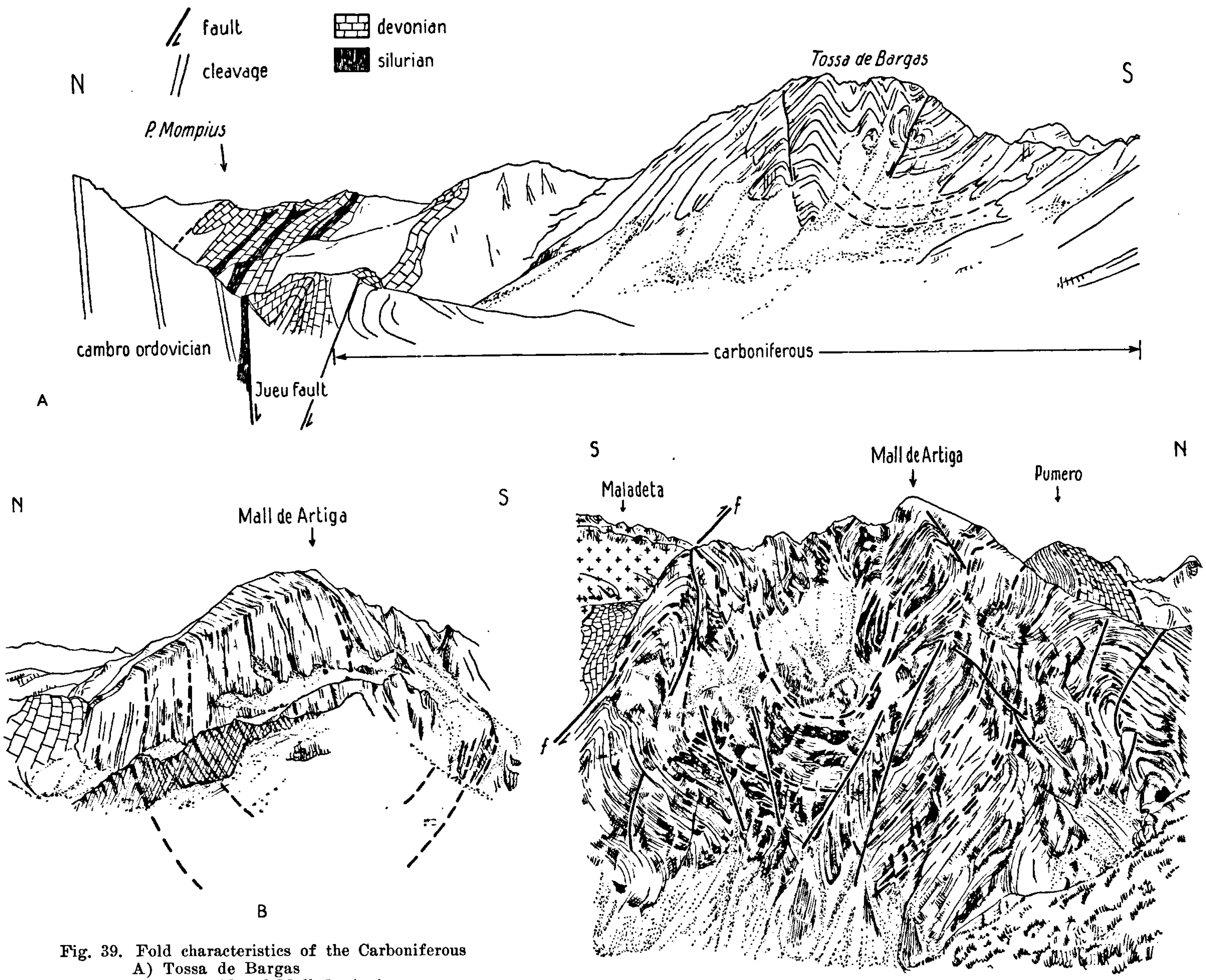


Fig. 39. Fold characteristics of the Carboniferous
A) Tossa de Bargas
B) West side of Mall de Artiga
C) East side of Mall de Artiga

G. LATE-CARBONIFEROUS INTRUSIVES

a. Dykes

Among the dykes there occur two important groups; the *quartz-porphyries* and the *hornblende-diorite-porphyries.*

The *quartz-porphyries* occur frequently in all Palaeozoic strata exposed on sheet 4. Sometimes the original structure can still be recognized, pointing to much altered porphyric rocks with phenocrysts of quartz, feldspar and dark minerals.

The majority of the quartz-porphyries occur in one elongated zone (Fig. 42). This dyke belt can be divided into 3 parts according to maroscopic aspect rather than to mineralogical composition.

Fig. 40. Cleavage developed in small Carboniferous folds in the upper part of the sequence exposed in the Els Neres

The dykes which occur in part A are light grey to light yellow-grey, fine-grained and consist of very hard rocks without signs of cleavage; joints, although common, are not well-developed. Black or brown-black aggregates of biotite not more than 5 mm long are characteristic of these dykes. Quartz forms 70 to 80 % of the matrix. Feldspar and biotite have been altered into sericite, zoisite and muscovite; calcite occurs rarely.

The dykes of part B are more or less transitional between A and C; they are also light-coloured, fine-grained hard rocks. Fracture cleavage, well-developed in the dykes of part C, is not as frequent. The characteristic dykes of part A with biotite as the only phenocrysts are absent. Phenocrysts are biotite, quartz, occasionally plagioclase, and pyrite. Phenocrysts and matrix are almost completely altered; calcite is more frequent than in part A.

The dykes of part C are characterized by a well-developed fracture cleavage running parallel to the fracture cleavage of the surrounding sandy slates; joints are better developed than in part A. They are dirty-white, light red-brown or orange-brown coloured dykes. The matrix contains con-

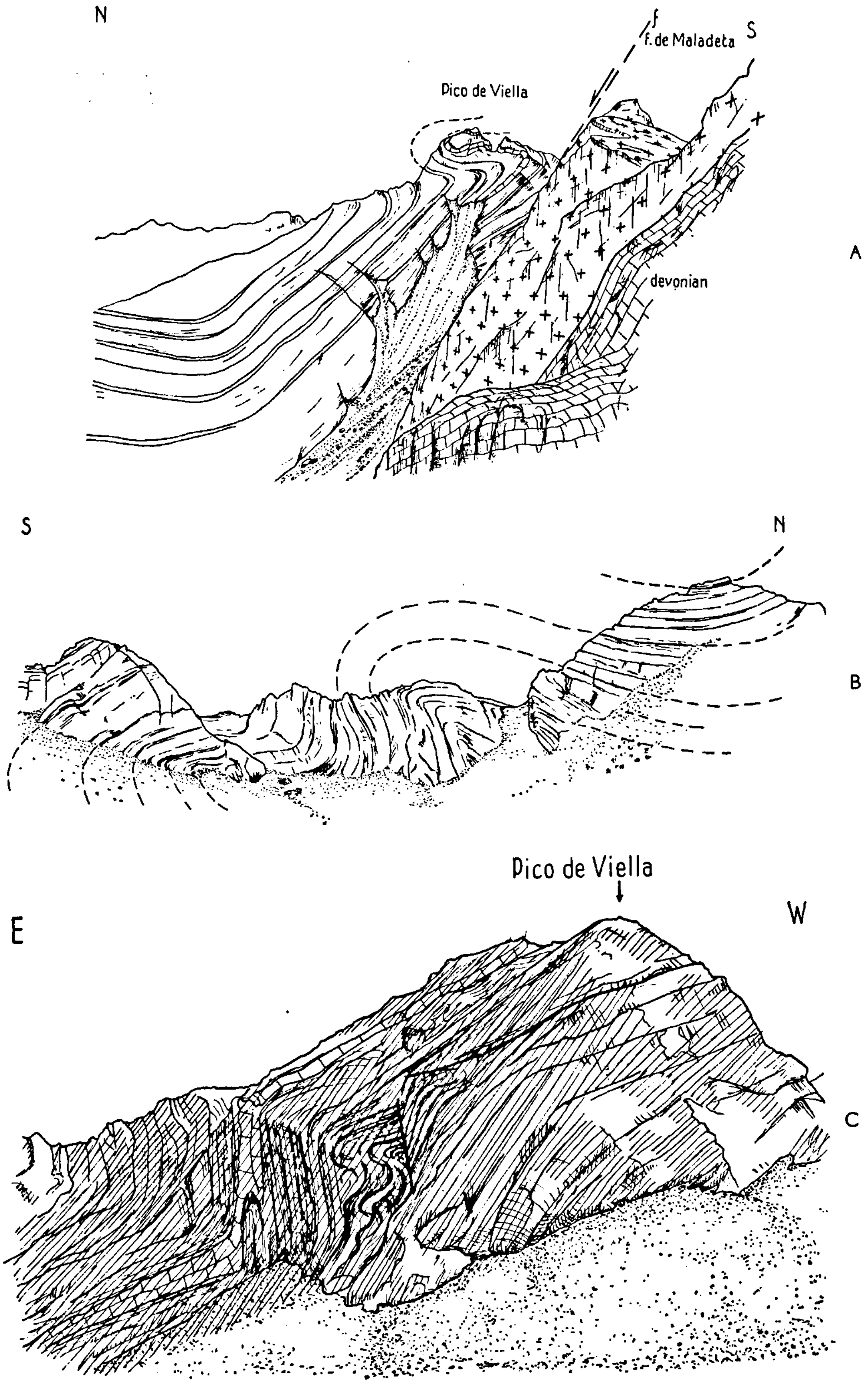


Fig. 41. A) Recumbent fold in the Carboniferous of the Pico de Viella, seen from the west. After photograph
B) Detail of the recumbent fold of the Pico de Viella, seen from the east. After photograph
C) Longitudinal section through the Carboniferous of the Pico de Viella. After photograph

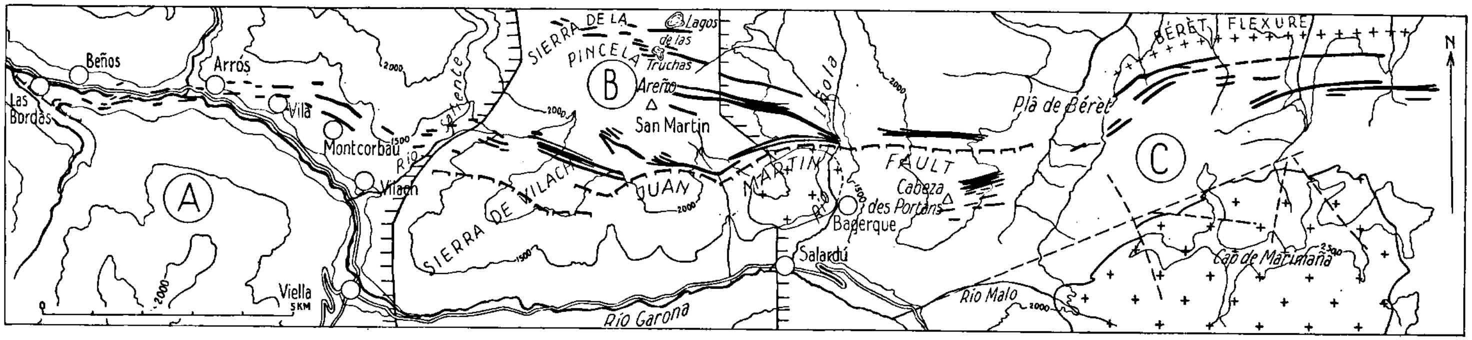


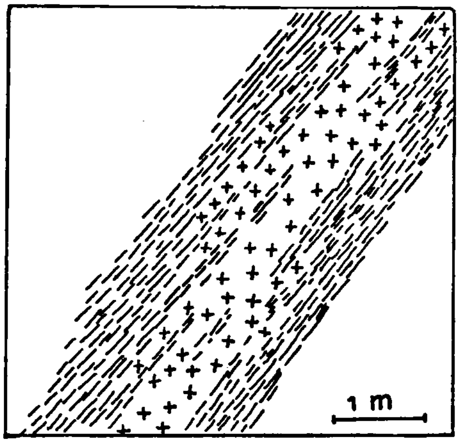
Fig. 42. The dyke belt of the Valle de Arán

siderably less quartz than the matrix of the dykes of part A, and is almost completely sericitized. Phenocrysts are quartz, biotite and albite, the latter strongly altered into sericite, chlorite and zoisite; calcite is common.

Generally the dykes are altered to a considerable degree, reaching a maximum in part C. Calcite has been found as an alteration product of feldspar, but has also been introduced in larger quantities from outside. The boundary with the host rock is always sharp; contact metamorphic aureoles round the dykes are missing. Cataclastic textures like cataclastic rims, broken quartz phenocrysts and wavy extinctions of quartz are common, especially in the dykes of part C; they generally show a schistose texture.

When the different parts are compared increasing amounts of sericite, chlorite and calcite and decreasing amounts of quartz can be noted; fracture cleavage becomes well-developed towards the east.

The quartz-porphyries of the Plâ de Bérèt penetrate wedge-like, "lit par lit" into the surrounding slates (Fig. 43). In one place these quartz-porphyries reach a thickness of 250 m, but they contain numerous xenoliths of slates.



after J. Snoep

Fig. 43. Quartz-porphyric dykes intruded parallel to the cleavage of the surrounding slate.

Nowhere could it clearly be proved that these dykes are folded together with the surrounding rock, although they occur parallel to the cleavage of the surrounding slates. In other places they occur both as concordant and cross-cutting bodies with regard to cleavage not bound to a fixed stratigraphic horizon.

The absence of fracture cleavage in part A is probably due to the amount of quartz (70—80 %) and the small percentage of sericite and chlorite, rather than to assumed intrusion later than the main folding.

Raguin (1946) described similar dykes from the Pic de Paragrano and Pic de Bulard (sheet one), which he considers to be rhyolites extruded during an Upper-Ordovician epirogenic movement. His opinion was based on their great thickness (30 m), extension (2 to 3 km) and occurrence on one stratigraphic level: the Upper-Ordovician. He also regards these dykes as folded together with the sediments. Absence of fracture cleavage, such as is common in the Cambro-Ordovician dykes of the Plâ de Bérèt, however, may indicate a post-tectonic intrusion.

Keyzer (1954) described more or less similar dykes from the Arize massif as intrusive. Also Snoep (1956) described the quartz-porphyries of the Plâ de Bérèt as intrusives.

Dykes similar to those previously described are found in the Cambro-Ordovician of the Central Anticline and in the Silurian of the Sierra Negra and Mail du Criq, as well as in the Devonian and Carboniferous.

Many fine-grained dykes of a porphyric character, closely resembling the quartz-porphyries, grey or reddish-grey, swarm out from the granite of Lys-Caillaouas (457/47) into the surrounding rocks. Phenocrysts are quartz, felspar, plagioclase and hornblende largely altered into sericite and chlorite. Hornblende, usually chloritizised, is common.

Hornblende-diorite-porphyries are known from the Puerto de la Bonaigua, a location west of the Rio Negro valley, the Pico de Entecada and from a locality about 600 m to the east of Cazarilh. Their colour is green-grey when found in calcareous deposits, otherwise dark-green or brown. These dykes have phenocrysts of hornblende, felspar and quartz; the Bonaigua dykes have brown hornblende (40 %), in those of the Entecada the hornblende is brown and green (50 %). Phenocrysts as well as matrix have been largely altered into chlorite, sericite, muscovite, calcite and quartz. The absence of calcite in the hornblende-diorite-porphyries of the Rio Negro is noteworthy because these dykes occur in limestone. Snoep (1956) described from the Puerto de la Bonaigua dykes, showing clearly "chilled margins".

Basic dykes are occasionally found in the granodiorites and in the eastern part of the Northern Anticline.

The origin of the dykes, especially the quartz-porphyries, is not claer. It is supposed that they are associated with the granodiorites because they are known to swarm out from the granodiorites of Lys-Caillaouas and of Bono *) and Barruera *), small stocks south of the Maladeta s.s. (Fig. 8). None of the other granodiorites, however, shows a direct connection between dykes and intrusive masses.

b. Granodiorites

Granodioritic batholiths intruded at the end of the Hercynian orogene. They comprise the extensive mass of the Maladeta s.l., the smaller intrusions of El Pruedo, Arties, Tredós, Salardu, Marimaña and a very small stock in the Sierra Negra. The granite of Lys-Caillaouas differs in some respects from the granodiorites as is shown in Table II.

The granodiorites, except the Salardu stock and the Sierra Negra stock, are surrounded by Devonian or Cambro-Ordovician (Marimaña) limestones. It appears as if the intruding masses had been stopped by these limestone horizons.

The metamorphic influence of the granodiorites apparently is rather small; pelitic sediments are converted into hornfelses and spotted slates with

*) Numerous dykes sometimes showing a well-developed fracture cleavage swarm out from both small stocks.

TABLE II

Composition and characteristics of the Biotite-Granodiorites and the Granite of Lys-Caillaouas.

Composition	Biotite-Granodiorites	Granite of Lys-Caillaouas
Quartz	23 %	25 %
Plagioclase	50 %	30 %
Potassium Felspar	2 %	30 %
Hornblende	3 %	—
Muscovite	—	small quantities
Biotite	18 %	12 %

Characteristics	Biotite-Granodiorites	Granite of Lys-Caillaouas
Dark inclusions	small-sized inclusions rich in biotite are common	Large floes of micaschists, sometimes hundreds of meters long.
Oriented minerals	Not observed	Potassium felspar phenocrysts show preferred orientation (flow?).
Metamorphic aureole	Small; spotted slates and marbles	Large; granite entirely situated in micaschists.
Limestone inclusions	Frequent; sometimes of great dimensions	Not observed.
Contact with surrounding sediments	Generally sharp; locally concordant	Generally sharp; locally concordant.
Aplites	Present	Common.
Basic dykes	Present	Common.
Pegmatites	Only near Lago Paderna	Common.
Granitic dykes	Only near Pico de Viella	Not observed.
Quartz porphyritic dykes	Present	Common.

andalusite porphyroblasts. Snoep (1956) also mentioned rare chiastolite porphyroblasts from the Cambro-Ordovician black slates north and west of the Marimaña. The limestones surrounding the granodiorites are marmorized; the boundary between granodiorite and limestone is often very sharp.

Calc-silicate rocks are frequent; the top of the Pico Salana is composed of epidote-hornfels. Calcium assimilation of the granodiorites is demonstrated by frequent calcium-rich minerals, mainly hornblende near their boundaries).

The pegmatites near the Lago Paderna (south of the Pico de Paderna) are rich in tourmaline and plumose mica. Metamorphic greywackes in this area may be extremely rich in muscovite.

Spotted slates with andalusite porphyroblasts are still found at a distance of 2½ km from the granite of Lys-Caillaouas. Approaching the granite we find alternating layers of andalusite, biotite and quartz. A broad zone of gneiss consisting of biotite, andalusite, quartz and a high percentage of muscovite borders the granite; plagioclase is completely missing. The large amount of muscovite is probably due to a late phase of metamorphism.

Chemical analyses of some dykes and granites are given in Table III.

TABLE III

Chemical analyses of some dykes and granites of the Valle de Arán.

	Hornblende Diorite	Hornblende granite	Quartz-Porphyrite	Lamprophyre	Quartz-Porphyrite	Muscovite Diorite
SiO_2	58.45	50.10	68.45	59.73	70.28	71.60
TiO_2	1.06	1.14	0.10	0.83	0.34	—
P_2O_5	0.52	0.28	0.29	0.47	0.28	—
Al_2O_3	16.27	14.44	15.62	16.48	14.09	14.40
Fe_2O_3	2.18	1.67	1.42	0.84	0.25	0.90
FeO	3.34	6.89	2.29	4.18	1.83	1.40
MnO	0.11	0.14	0.05	0.08	0.04	—
MgO	4.10	9.71	0.21	3.14	0.52	0.40
CaO	5.16	7.60	3.49	5.92	1.63	1.60
Na_2O	3.58	2.14	3.34	2.79	2.40	3.50
K_2O	2.17	2.46	3.53	2.64	5.48	5.00
H_2O+	2.75	2.63	0.99	2.09	2.46	—
H_2O-	0.34	0.57	0.09	0.29	0.31	—
CO_2	—	0.35	—	0.72	—	—
	100.03	100.12	99.87	100.20	99.91	98.80
	Rio Negro	Puerto de la Bonaigua	Las Bordas	Maladeta s.s.	Pico	Bosost

Analyst: Dr. C. M. de Sitter-Koomans.

c. Tectonics of the Intrusives

Signs of tectonisation, except jointing and faulting, are not found in the batholiths. They have been intruded after the last folding phase of the Hercynian orogene.

The granodiorites of Arties, Tredós and the Maladeta s.l. greatly domed up the surrounding limestones. This effect is much less marked in the Marimaña area and seems absent around the Salardu stock, which presumably caused faulting only (Juan Martin fault).

All granodiorites have "shouldered aside" their surrounding sediments. At the boundaries of the Marimaña, Tredós and Salardu granodiorites this "shouldering aside" is clearly shown in the arrangement of lineations parallel to their boundaries.

The "shouldering aside" effect of the Maladeta s.l. is not as clear. In the case of this large intrusion, however, it is reasonable to suppose similar effects. Structurally the influence of the Maladeta s.l. must have been considerable but has not yet been fully understood.

The strong compression of the Carboniferous, the sudden narrowing and pronounced folding of the Central Anticline, accompanied by large faults, might have been caused by the intruding Maladeta batholith. The "shouldering aside" effect to the north could be responsible for the recumbent fold of the Pico de Viella and for the accompanying complications.

The granite of Lys-Caillaouas differs somewhat in character from the granodiorites.

Most of the quartz-porphyric dykes probably intruded before the emplacement of the granodiorites. Obviously their intrusion must have taken place after the main folding phase in view of the absence of slaty cleavage in the dykes and their injection parallel to the cleavage of the surrounding pelites. But after the intrusion of the dykes a second period of tectonical activity, less intensive, must have taken place. This caused the sometimes well-developed fracture cleavage in part of the dykes and the clastic deformation and recrystallisation of its minerals.

H. OBSERVATIONS ON CLEAVAGE AND RELATED FEATURES

It has been noted that cleavage planes in pelitic sediments run strictly parallel to the axial plane of the fold. Deviation of cleavage planes results

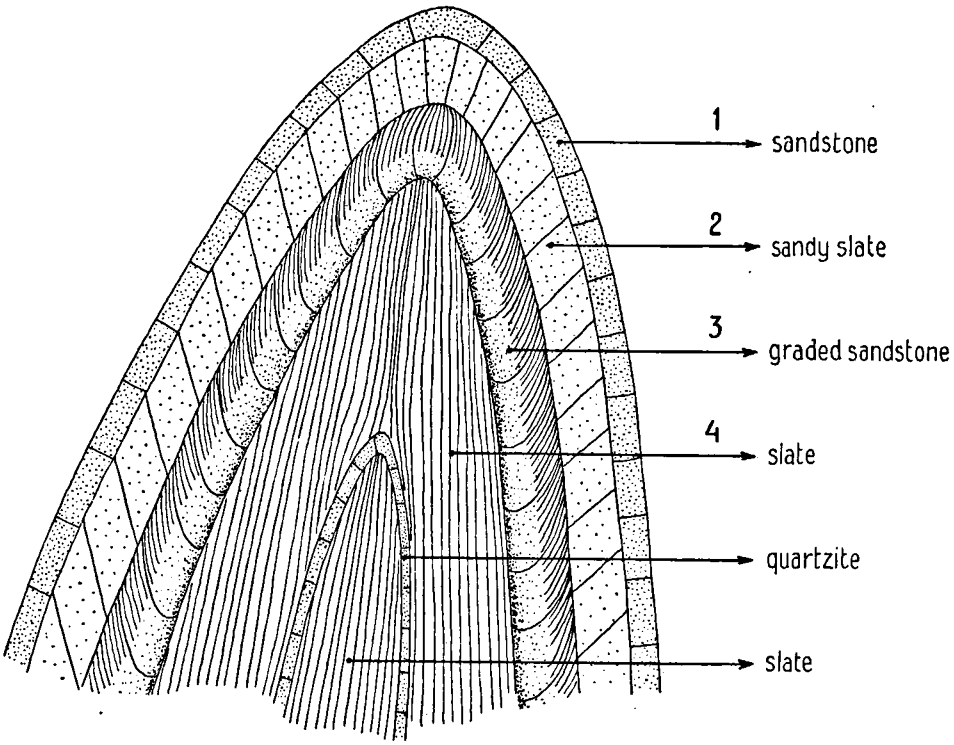


Fig. 44. Development of cleavage in beds of different lithology:
1) Rotational joints in sandstones and quartzites
2) Deviated cleavage in sandy slates
3) Curved cleavage in graded beds
4) Slaty cleavage in pelitic rocks

from differences in lithology (Fig. 44). Fan-shaped cleavage (Fig. 44, No. 2) is generally found in sandy slates; Curved cleavage, resulting from gradual changes in lithology, can best be observed in graded beds (Fig. 44, No. 3). If the boundary between slate and sandy slate is not sharp the cleavage planes may also be curved in the transition zone. Curving of cleavage planes, occasionally observed in thin incompetent layers situated between competent beds, however, is produced by drag along the competent layers.

Great use has been made of cleavage-bedding intersections in finding the direction of younging with the help of such differences in cleavage direction (Wilson, 1946, p. 269—270). As slaty cleavage and bedding in steep isoclinal folds often run nearly parallel, better results were obtained in using deviated cleavage and the curved cleavage of graded beds. Bottom-top criteria of the graded beds indicated that recumbent folding does not occur, although the folds generally are slightly overturned.

Cleavage characteristics largely depend on the lithology (Scholtz, 1930, 1932); their connection has been studied in the field and in thin sections.

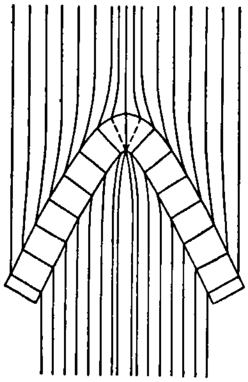


Fig. 45. The development of outward bend cleavage (outer anticlinal hinge) and inward bend cleavage (inner anticlinal hinge)

The spacing of the cleavage clearly depends on the grain size; the coarser the grain, the wider the spacing. Cleavage planes, usually smooth, become irregular on entering coarser material.

The gradual deviation of cleavage in graded beds depends on grain size and increasing grain percentage. Cleavage in different lithologies generally deviates to the side of greater resistance (Fig. 44, No. 2). There is a connection between the deviation and the angle of incidence; it increases as the angle decreases. The maximum deviation from the axial plane is 45°.

In Devonian sandstones, in which the grains are equally distributed in the matrix, development of cleavage planes becomes apparent in thin sections when the matrix percentage exceeds 40 %. Although cleavage planes are not observed in sandy beds of lower matrix percentage (about 30 %), the grains still might give the impression of being oriented. It should be noted, however, that these percentages largely depend on the intensity of the cleavage folding and are only valid for a restricted region. But it seems fairly certain that the development of cleavage also depends on the lithology.

Cleavage in pure sandstones or quartzites does not occur in our region. Only in the anticlinal or synclinal hinges of folded sandstones in which cleavage is not observed does a coarsely-spaced (fracture?) cleavage become apparent. In overturned parasitic folds in thin sandstone layers, concentration of cleavage planes in the overturned flank might be able to cut through the sandstone.

In small-scale folds composed of alternating thin competent and incompetent layers cleavage planes generally converge upwards. But large-scale converging of cleavage is also evident in the Central Anticline near the Pico de Escaleta (Fig. 19, sections 16 & 17).

In isoclinal folds outward bending of cleavage planes around the anticlinal hinge, and somewhat less pronounced inward bending of cleavage planes in the synclinal hinge of competent beds is general (Fig. 45). This phenomenon might indicate cleavage development in an early stage of the folding. Each cleavage plane corresponds to a fixed position in the competent bed. Thus shortening, equally distributed in cleavage folding, results in deviation of cleavage planes around the hinge of the concentrically-folded competent layer (Fig. 45). This would not occur, or would hardly be noticed, if the cleavage had developed at a relatively late stage of folding.

On the other hand, deviation of cleavage towards the side of greater resistance in beds of different lithology might indicate that some initial folding took place before cleavage developed. Scholtz (1932, p. 304) assumed cleavage development at a very early stage of the folding because of the fan-shaped cleavage in sandy beds. Mead (1940), however, supposes a relatively late cleavage development. E. Cloos (1947) clearly demonstrated that visible cleavage already developed with a shortening of 20 %.

As most of the sedimentary rocks in the Valle de Arán are found in very low-grade metamorphic areas (outside the Bosost area as well as outside any contact-metamorphic aureole), complete recrystallization of the pelitic sediments is mainly the result of the cleavage process.

According to De Sitter (1956-a, p. 67), cleavage planes enclose narrow slices (microlithons) which are compressed and parallel extended, slipping over each other. Since cleavage might be indicated as a structurally controled rupture (Fairbairn, 1949, p. 76) under high confining pressure, shearing is the most important factor for the formation of new, cleavable minerals like mica (sericite) and chlorite in chemically unstable bodies (slates and impure limestones). Collapse under increasing strain is shown by the development of shear planes perpendicular to the direction of stress (Scholtz, 1930). The resulting energy, converted into heat, produces secondary minerals (G. F. Becker, 1893, p. 13). Others, however, suppose that the process is purely mechanical; detrital grains are turned and flattened in planes perpendicular to the main stress (Harker, 1885, Goguel, 1945).

Minor structures associated with cleavage folding are boudins, parasitic folds and mullions (p. 199). Such structures are found only in relatively competent layers situated between incompetent rocks. Small-scale boudins and parasitic folds in competent layers have occasionally been observed in intercalated limestones and, to a lesser extent, in sandstones of the Central Anticline.

I. MICRO-STRUCTURES

a. Lineations

On the assumption that cleavage is syn-tectonic, cleavage planes should run more or less parallel to the axial planes of the folds. Intersection of bedding and cleavage planes therefore gives a lineation parallel to the structural plunge.

In sandy beds, where the bedding is clear, lineations can best be measured on the bedding plane. In slates minor lithological changes may be expressed by slight differences in colour, which can be used to measure lineations on the cleavage planes.

The intersection of cleavage and bedding shows on the bedding plane as closely-spaced, fine striae, each corresponding to a cleavage plane. The spacing depends on the grain size of the sediments. Bedding and cleavage

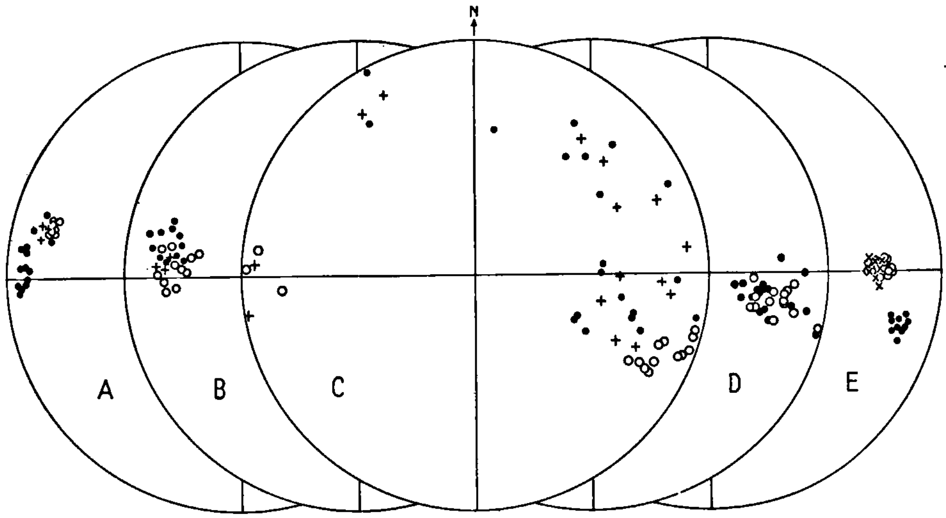


Fig. 46.

A. Lineations measured in one single fold (intersection slate-sandstone)
B. Lineations measured in one single fold (intersection sandy slate-sandstone)
open circles: lineations north of the anticlinal hinge
solid id.: lineations south of the anticlinal hinge
crosses: plunge of fold axes measured on different intervals
C. Micro-folds in Carboniferous limestones
open circles: plunge of larger fold axes in the Carboniferous
solid id.: plunge of anticlinal axes of micro-folds
crosses: plunge of synclinal axes of micro-folds
D. Parallel arrangement of lineations and fold axes, Corbison area
open circles: fold axes
solid id.: lineations
E. Parallel arrangement of mullions
open circles: Las Bordas area
solid id.: Casau
crosses: Between Las Bordas and Viella

might often run roughly parallel in isoclinal folds, but a small angle between them is quite sufficient to cause lineation.

In folds in contrasting Devonian sediments the connection between fold-axes and lineations has been studied, and in all cases the plunge and direction of plunge were found to be roughly equivalent. Data for one of these folds have been plotted in a Schmidt's net (Fig. 46-A). The plunge of this fold was measured at different intervals.

Cleavage planes in folded sandy pelites generally deviate from the

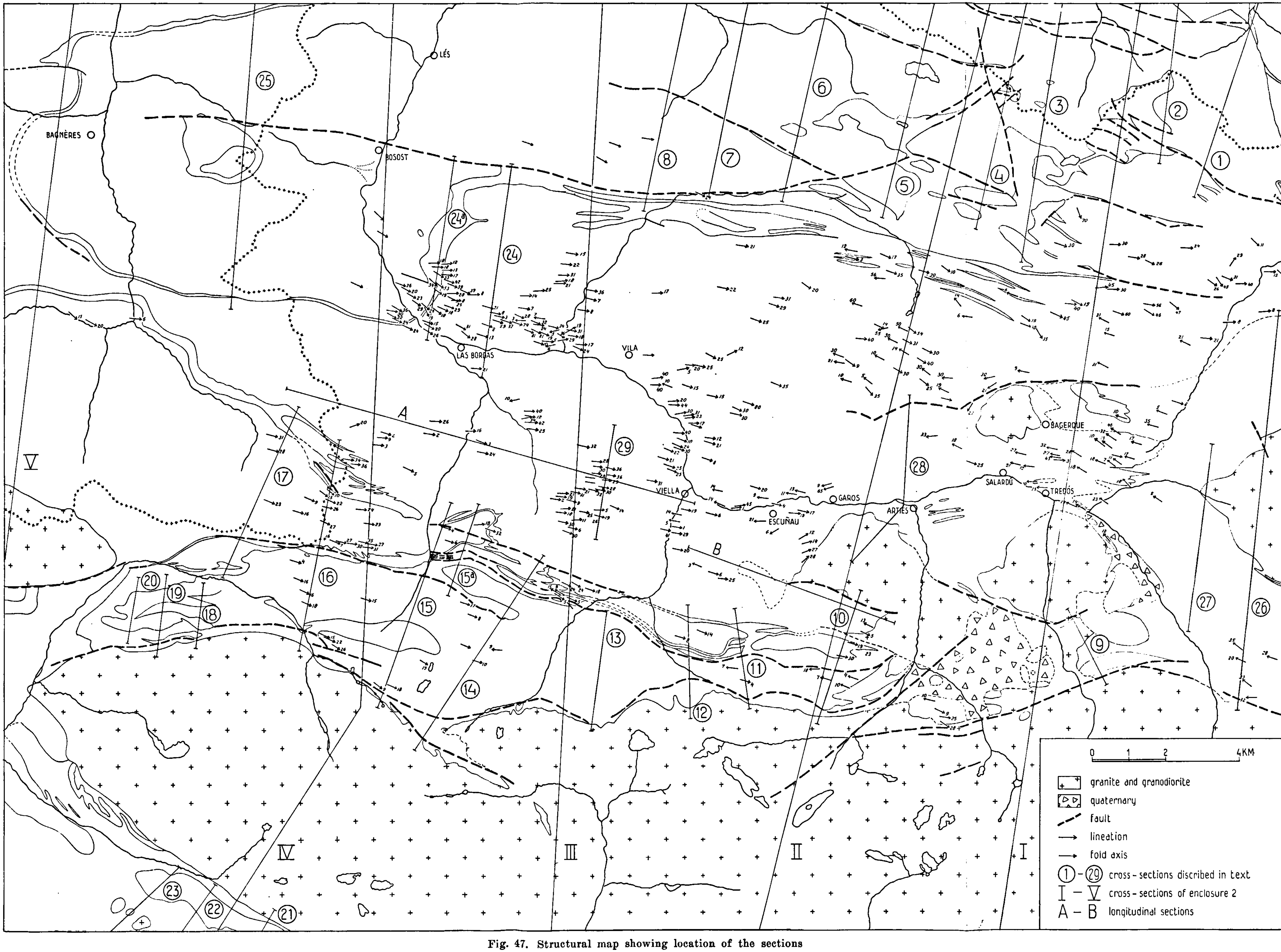


Fig. 47. Structural map showing location of the sections

orientation of the slaty cleavage towards the side of the greater resistance (Fig. 45-2). But here also the plunge of lineation and the fold axis are more or less parallel (Fig. 46-B). Thus for individual folds the syn-tectonic origin of the cleavage has been demonstrated. The parallel arrangement of fold axes and lineations is clear (Fig. 46-D) also for a larger area, such as the east slope of the Corbison (Fig. 37, section 29).

In the Las Bordas sandstones exceptionally steep, but equivalent plunge of lineations and fold axes are frequent. Plunge reversals of the same trend were also occasionally observed.

Parallelism of lineation and fold axis may therefore be assumed for the whole area under discussion. An exception must be made for small-

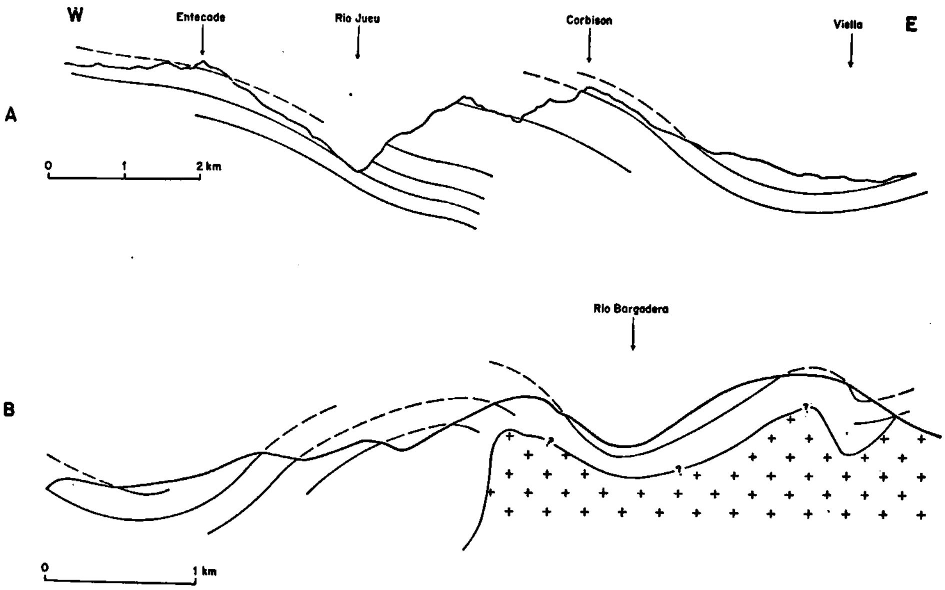


Fig. 48. Longitudinal sections through the Devonian

scale folds in metamorphic limestones or in limestones showing flow structures. Plunges have been measured in a strongly micro-folded, thin Carboniferous limestone over a distance of more than 1 km. As shown in Fig. 46-C these plunges are widely scattered. Many fold axes and lineations were measured, especially in the Devonian of the Valle de Arán. Plunges, although variable, in general change gradually along the fold axes as shown in Fig. 47. A longitudinal section through the Devonian of the Pico de Entecada to the Rio Valarties, partly constructed from fold axes and lineations, is given in Fig. 48.

The structures south of the Maladeta s.s., those of the isolated Carboniferous of the Paderna area and those in the extreme north-west, plunge to the west. The generally eastward-plunging structures in the west and the westward-plunging structures in the east meet in the Areño area. Alternation of strongly east and west-plunging structures is common in this plunge depression, the two opposing plunge directions interfingering here.

b. Fracture Cleavage and folded Cleavage

Fracture cleavage (secondary cleavage), characterized by coarse spacing and irregular cleavage planes, intersects the slaty cleavage (primary cleavage).

In sandy series of the Plâ de Bérèt area coarse fracture cleavage of about 1 cm spacing nearly obliterates a fine-spaced cleavage of an other orientation. This fine-spaced cleavage runs parallel to the cleavage of the underlying pelites. A coarse fracture cleavage was also noted in surrounding quartzporphyric dykes, but no primary cleavage. The plane of the fracture cleavage is much wider than the plane of the slaty cleavage. New platy minerals were not formed, but the filling of the fracture with quartz or calcite is typical of fracture cleavage.

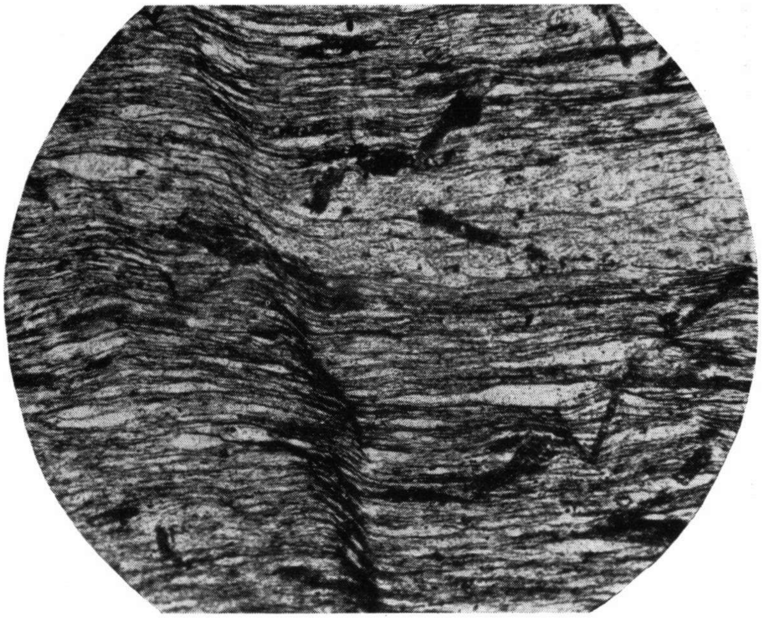


Fig. 49. Folded cleavage in Entecada slates. 67 ×

When thin sections are examined neither dilatation nor shortening is observed in the slices enclosed by the fracture cleavage planes; the movement of the slices often alternates, causing curving of the cleavage planes. In thin sections of slate as well as in sandy slates and in sandstones folded cleavage planes are occasionally observed (Fig. 49), sometimes accompanied by fracture cleavage. Folded cleavage and fracture cleavage seem closely associated. The folding of the cleavage planes might be very pronounced grading into continuous chevron folding or even knicking.

These phenomena are much more pronounced in the Cambro-Ordovician than in the Devonian. In thin sections of Cambro-Ordovician micro-conglomerates every small slate pocket shows the phenomena described above.

Zwart (1959) described the occurrence of chloritoid in the Valle de Arán in relation to fracture cleavage and folded cleavage. Scholtz (1930)

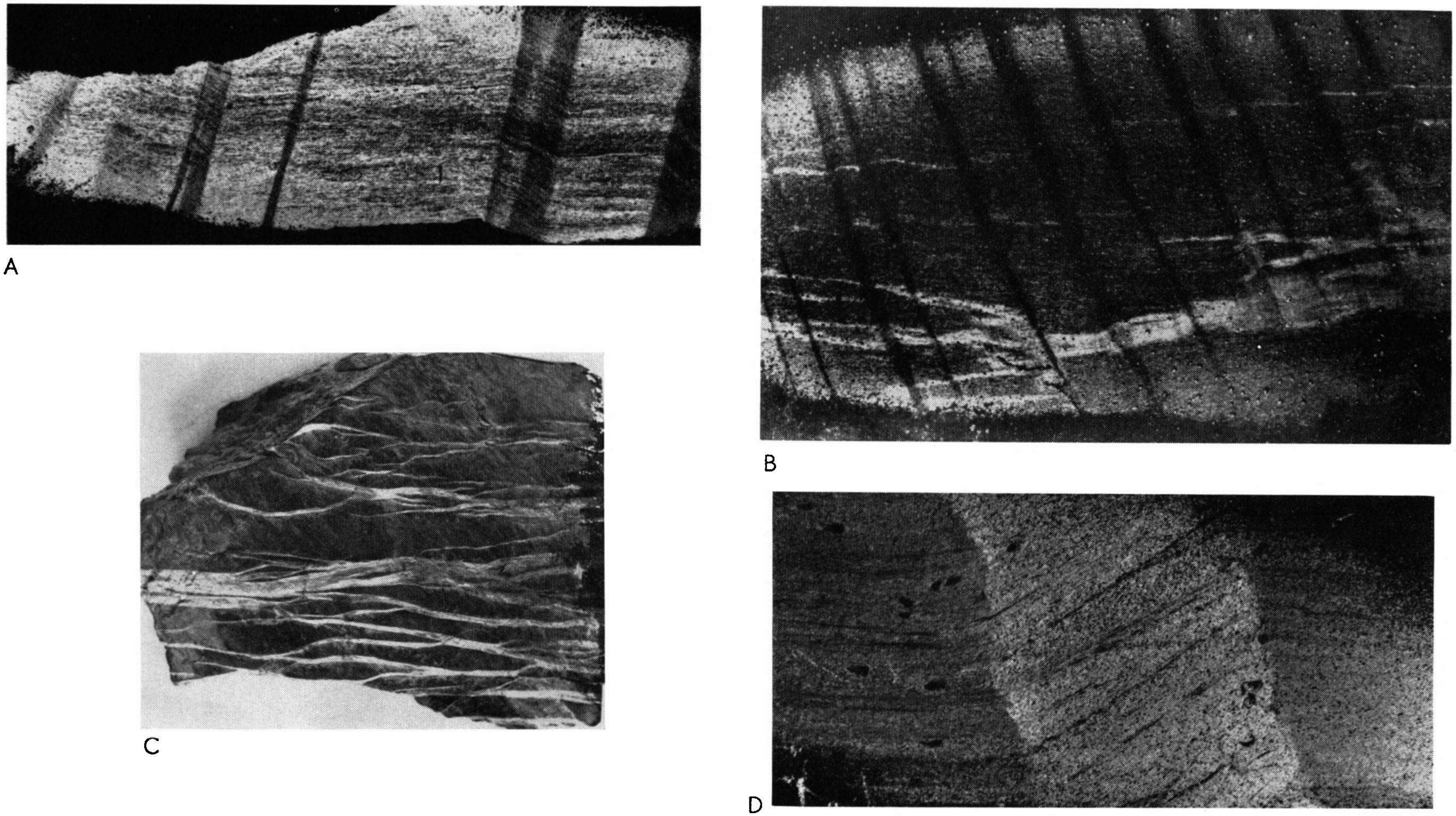


Fig. 50.
A) Angular knick-planes in Devonian slates, north-east of Viella. 3½ ×
B) Approaching and slightly curved knick-planes in Carboniferous slates, south of Prat. 3½ ×
C) Irregular knick-planes in Devonian slates, north-east of Viella. 2 ×
D) Slightly rounded knick-planes in Devonian slates, north-east of Viella. 3½ ×

described fracture cleavage intersecting slaty cleavage; folding, cleavage and fracture cleavage are due to the same cause. The spacing of the fracture cleavage depends entirely on the lithology. Kienow (1934) considers fracture cleavage (and associated folding of cleavage planes) as a direct continuation of folding and cleavage mechanism and of the same origin. According to Wilson (1946) fracture cleavage is the result of a purely mechanical process, the spacing depends largely on rock characteristics. Continuous stress may wrinkle the cleavage planes and the slates may undergo fracture cleavage.

c. *Knicked Cleavage*

In the Cambro-Ordovician of the Central Pyrenees knicked cleavage sometimes occurs, especially in fine grained slates. This phenomenon is also locally abundant in Devonian slates and even in fine-grained Carboniferous slates.

Knick planes generally intersect the slaty cleavage of the country rock in parallel pairs, thus forming zig-zag flexures or knick zones. The distance between the knick planes of one pair may vary from less than a millimeter (as may be observed in thin sections) to about 10 cm. Zandvliet (1960) described much larger intervals in the Cambro-Ordovician of the river Pallaresa, east of our area (sheet 5). Generally the distance between two pairs is greater than the distance between the knick planes of one pair. Although mostly developed as straight lines (Fig. 50-A), they may occasionally be curved (Fig. 50-B) or even be irregular (Fig. 50-C). The planes sometimes gradually approach each other and die out (Fig. 50-D), forming lenses of knicked cleavage of widely-varying dimensions. Intersection of knick planes or knick zones has never been observed; approaching knick zones often continue as a single zone.

Geometrical data about knicked cleavage have been assembled in Fig. 51. Knick planes bisect the angle β between the two cleavage orientations. This angle may vary from approximately 160° to about 65°. If the angle is small, surface creep may have had some influence. It is not known whether, before knicking took place, the cleavage was parallel to one or other of the present cleavage orientations (Fig. 52-I), or whether it was intermediate (Fig. 52-II). Only in case I does offset occur, reaching a maximum when $\beta = 90°$. Observations in Devonian slates strongly support knicking of cleavage according to case I; the dip of the cleavage, however, may vary within relatively short distances. The knicking is always in the dip direction of the cleavage.

Knicking caused slip on the cleavage planes (already potential shear planes), increasing as angle β decreases. Slip along knick planes might take place in the rare instances where they become fault planes. In order to produce angular knicking, equal slip along all cleavage planes must have taken place (Fig. 53-A). When, on the other hand, only a few of the cleavage planes take most of the slip, a disturbed zone develops along the knick plane (Fig. 53-B). Such broken knick zones strongly resemble jointing. Angular knicking is more common; the knick points are generally sharp but some rounding has occasionally been observed in thin sections (Fig. 50-D).

Knicked cleavage is restricted to fine-grained pelitic rocks with strongly-developed cleavage. But even in such rocks knicking is not general. Cleavage and the knicking associated with it is apparently better developed

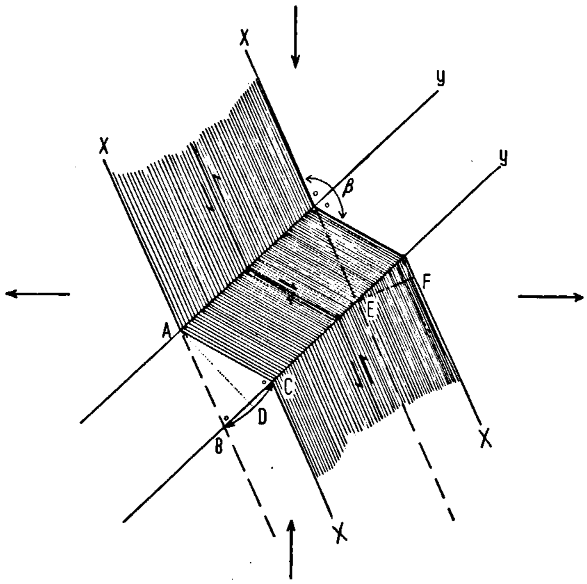


Fig. 51. Geometry of the mechanism of cleavage knicking

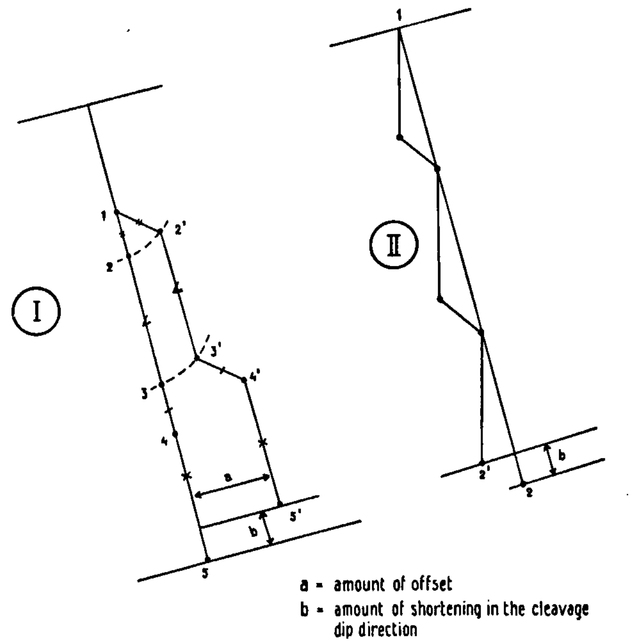


Fig. 52. Possibilities of cleavage knicking
I. Parallel to the longest side of the knicked cleavage
II. Intermediate

in the Cambro-Ordovician than in the Devonian or the Carboniferous. As compression of the rock is not to be expected, knicking of cleavage must result in a change in size. Contraction takes place in the dip direction of the cleavage planes and consequently dilatation perpendicular to the cleavage planes. Thus the direction in which contraction or dilatation occurs depends entirely on the orientation of the cleavage planes.

Deformative stress results in shortening, whereas tension results in dilatation. As the cleavage is generally sub-vertical, knicking of cleavage

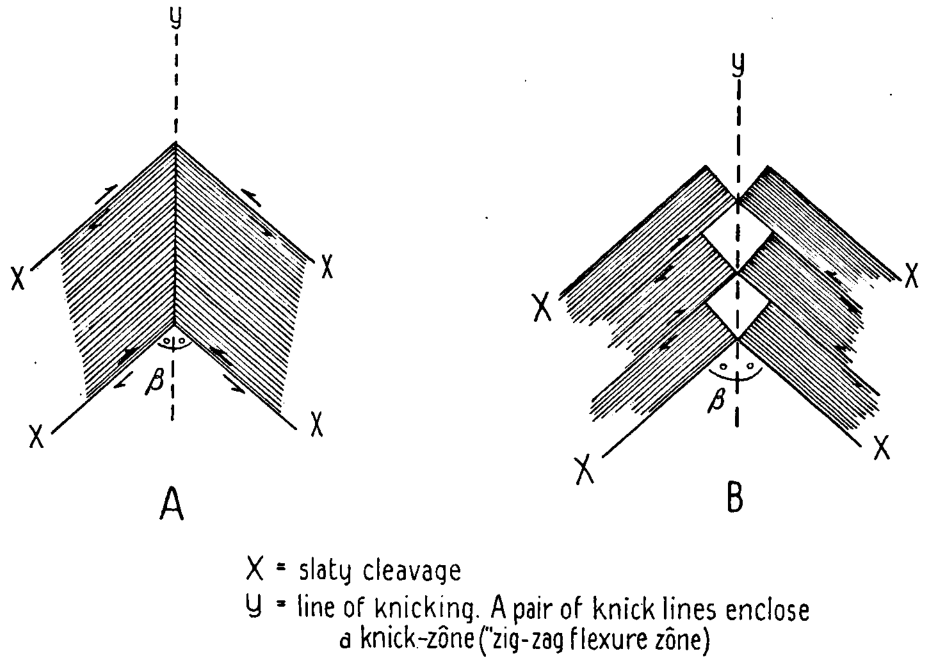


Fig. 53. Development of angular and of disturbed knicking of cleavage

results in dilatation in the horizontal plane. In other words, the knicking of cleavage is due to tension.

It seems reasonable to assume cleavage development in an orogene as a result of tangential forces. Cleavage planes therefore should originally have a more or less vertical orientation. The fan shape of the cleavage in the Pyrenees (Fourmarier, 1951) presumably originated from post-orogenic uplift. The resulting tension perpendicular to the fan-shaped cleavage and the force of gravity might account for the origin of knicked cleavage. This phenomenon presumably originated at the end of a post-orogenic Hercynian uplift.

Knicked cleavage is common and known from the Palaeozoic of Devonshire and Cornwall for instance and from the Rheinische Schiefergebirge. It has been described by several German authors (Born, 1929, p. 382, Kienow, 1934, p. 70, M. and R. Teichmuller, 1954, p. 266—267). According to Schenk (1954, p. 364) the shortening in knicked cleavage may be as high as 10 %. Hoeppener (1955, p. 34—35) and Simpson (1940, p. 34—35) described a form of knicked cleavage which is different from the knicked cleavage found in the Pyrenees (Fig. 54); knicked cleavage clearly associated with faulting is due to horizontal movements (sliding tensions). The mechanism, however, is the same as described in the case of the Pyrenean examples. Kienow (1951, p. 41) experimentally produced more or less similar features in plastic Tertiary clays. Simpson (1940) and Engels (1959, p. 76) mentioned the occurrence of intersecting zones of knicked cleavage.

d. *Mullions*

Mullions occur in the Las Bordas sandstones in those areas where relatively thick, well-layered sandstones and quartzites predominate. They are frequent south of Las Bordas (468/48), and are occasionally found along

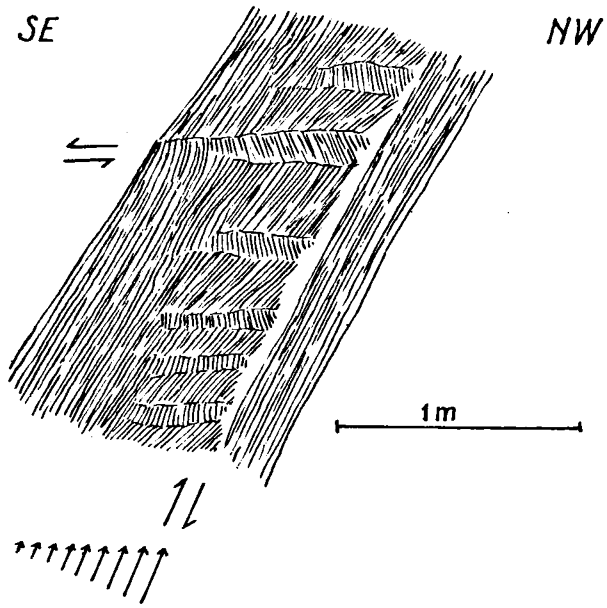


Fig. 54. Development of knicked cleavage as result of faulting (after Hoeppener, 1955)

the road from Las Bordas to Viella and near the village of Cassau (4°.28′/42°.42′).

Mullions are cylindrical columns of sandstone or quartzite. They often show striae parallel to their dip direction. They do not seem to have any connection with tight isoclinal folds, pinched-off crests of small isolated folds or even corrugated bedding planes.

South of Las Bordas the majority result from the intersection of bedding and coarse fracture cleavage (Fig. 55). Near Cassau their formation seems associated with boudinage. The origin of other mullions observed is less clear; they are probably associated with partly-developed parasitic folds, as they still show some of their features.

The plunge of the mullions is in accordance with those of other structures and the lineations. Strictly parallel arrangement is one of their main features, as is shown in Fig. 46-E.

Mullions are generally found in strongly-folded areas. Owing to their arrangement parallel to structures they seem closely associated with cleavage folding (Valle de Arán), although not all mullions observed are cleavage mullions in the sense of Wilson (1953).

Fig.55. Cleavage mullions of Las Bordas

Some authors regard mullions as a result of faulting (Gregory, 1914, p. 289, Leith, 1923, p. 100, Fairbairn, 1949, p. 178), as thrust accompaniments of folding (Read, 1926, p. 121, Philips, 1937, p. 581), or merely as the result of strong folding (Wilson, 1953, De Sitter, 1956-a, Schmidt, 1957). Wilson distinguished bedding or fold mullions, cleavage mullions and irregular mullions; De Sitter supposes also a close connection between certain mullions and boudins.

e. *Joints*

Macro-joints or cracks are particularly well-developed in the granodiorites and in the Northern Anticline. These joints show up on aerial photographs as straight or slightly-curved pronounced lines. Smaller joints are well-developed in the sedimentary rocks, and even in slates. Joints often run parallel to the broken knick planes of knicked cleavage.

Closely-spaced jointing has been observed in the Las Bordas sandstones north-west of the Mompius (Fig. 56). These sandstones form a gentle anticline in the core of a larger eastward-plunging syncline. The joints dip at 60°—70° in the direction S 48° E; they remain parallel for a long distance.

The age of the macro-joints is unknown. Their vertical position and independence of structures points to development after the main folding phase.

Mineralisation along macro-joints is common in the Northern Anticline. As this mineralisation is of Hercynian age, it seems fairly certain that the joints were formed in late-Hercynian times.

Fig. 56. Close jointing in sandstones south of the Pico Corbison. The joints near the point of the pencil are only partly developed

J. CHRONOLOGY AND CHARACTERISTICS OF STRUCTURAL EVENTS

In each of the large structural units of our region the type of folding is different for each large stratigraphic unit (Fig. 57). The Cambro-Ordovician occurs in large dome-like structures, the Devonian in medium sized isoclinal folds, the Carboniferous in large concentric folds. As a result of the incompetent character of the Silurian the strong compression of the Devonian folds has another character than the Cambro-Ordovician folds. The Devonian basal limestone penetrates deeply into the synclinal cores (Entecada anticline); in the upper synclinal regions many complications arise. Cleavage is apparently better developed in the older formations and in the pelites.

The cleavage folding developed at an early stage, probably preceded by weak concentric folding with some longitudinal imbrication or reversed faulting of minor importance in the competent units.

Intrusion of quartzporphyric dykes probably preceded the late Hercynian granite intrusions as the dykes often show fracture cleavage of some sort. Of the granite intrusions the Lys-Caillaouas intrusion seems to be the oldest one, as it shows some effects of tectonisation.

A late phase of the Hercynian orogenesis was characterized by longitudinal faults associated with a general arching of the whole axial zone. These faults were reactivated during Alpine movements like the structural depression north of the Maladeta massif.

The knicking of cleavage is certainly a late Hercynian phenomenon. It is a dilatation process, possibly associated with the faulting phase which supposedly preceded the granodiorite intrusions.

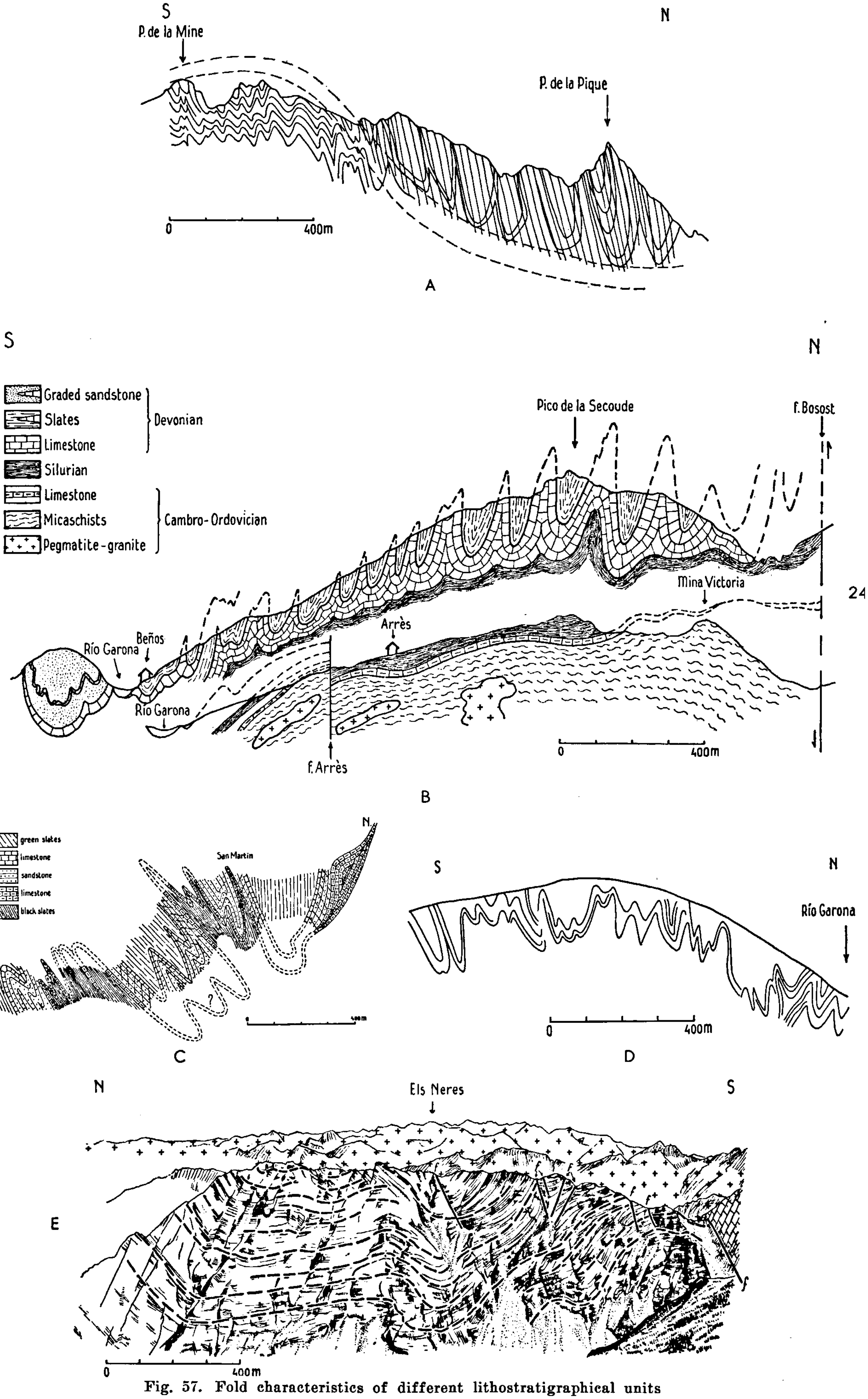


Fig. 57. Fold characteristics of different lithostratigraphical units

CHAPTER III

SEDIMENTATION

A. UPPER PALAEOZOIC SEDIMENT ASSOCIATIONS

In the upper Palaeozonic sandy sediments of the Valle de Arán three different litho-facies associations are distinguished, viz. an orthoquartzite-limestone association (Pettijohn, 1957, p. 611—615), a turbidite association and a molasse *) association (Pettijohn, 1957, p. 618—622).

a. *Orthoquartzite-limestone association*

This association is formed by the Entecada limestones and the overlying Las Bordas orthoquartzites. Orthoquartzite and limestone are both pure lithological units with an interposed variable horizon of thin slates.

The orthoquartzites consist of rounded quartz grains, which are highly cemented. Rounding is partly obscured by secondary growth. Feldspar occurs only in minor quantities. A photograph of a representative orthoquartzite sample made after a thin section is given in Fig. 58-B. The medium grain size of the orthoquartzites is 125 micron, the coefficient of sorting 2.59, the coefficient of skewness 1.54 and the kurtosis 0.37 (Pettijohn, 1957, p. 38).

The thickness of the orthoquartzites is persistent in a north-south direction, but decreases towards the east where they split up into several thin beds intercalated with slates. Towards the west they grade into slates in a relatively short distance.

Southwest of the Maladeta s.s., outside our area, a similar Devonian orthoquartzite-dolomite association occurs showing the same characteristics of persistency and thinning.

The sand laminae in the transition zone from orthoquartzite to slate in the west are highly broken up (Fig. 59). No burrowings or other sedimentary structures have been observed. It is unlikely that slumping, compaction or severe tectonization caused the mingling of sand and clay. We rather tend to the view that an initial, barely consolidated littoral or shallow marine deposit of alternating thin sand and clay laminae was ripped up by wave action.

b. *Turbidite association*

The Las Bordas formation, except the orthoquartzites and their transition zone, is recognized as a turbidite association owing to characteristic sedimentary structures of which grading is the most prominent (Ten Haaf, 1956, 1959a, b, Kuenen, 1953, Kuenen & Migliorini, 1950).

In the field, zones of different characteristics could be recognized in the turbidite association (Fig. 60). Transition from one zone to another was difficult to establish, so that the boundaries shown in Fig. 60 are only approximate. The differences are based on composition, predominance

*) Molasse is here used in the sense of a (paralic) sediment association and not as a tectofacies.

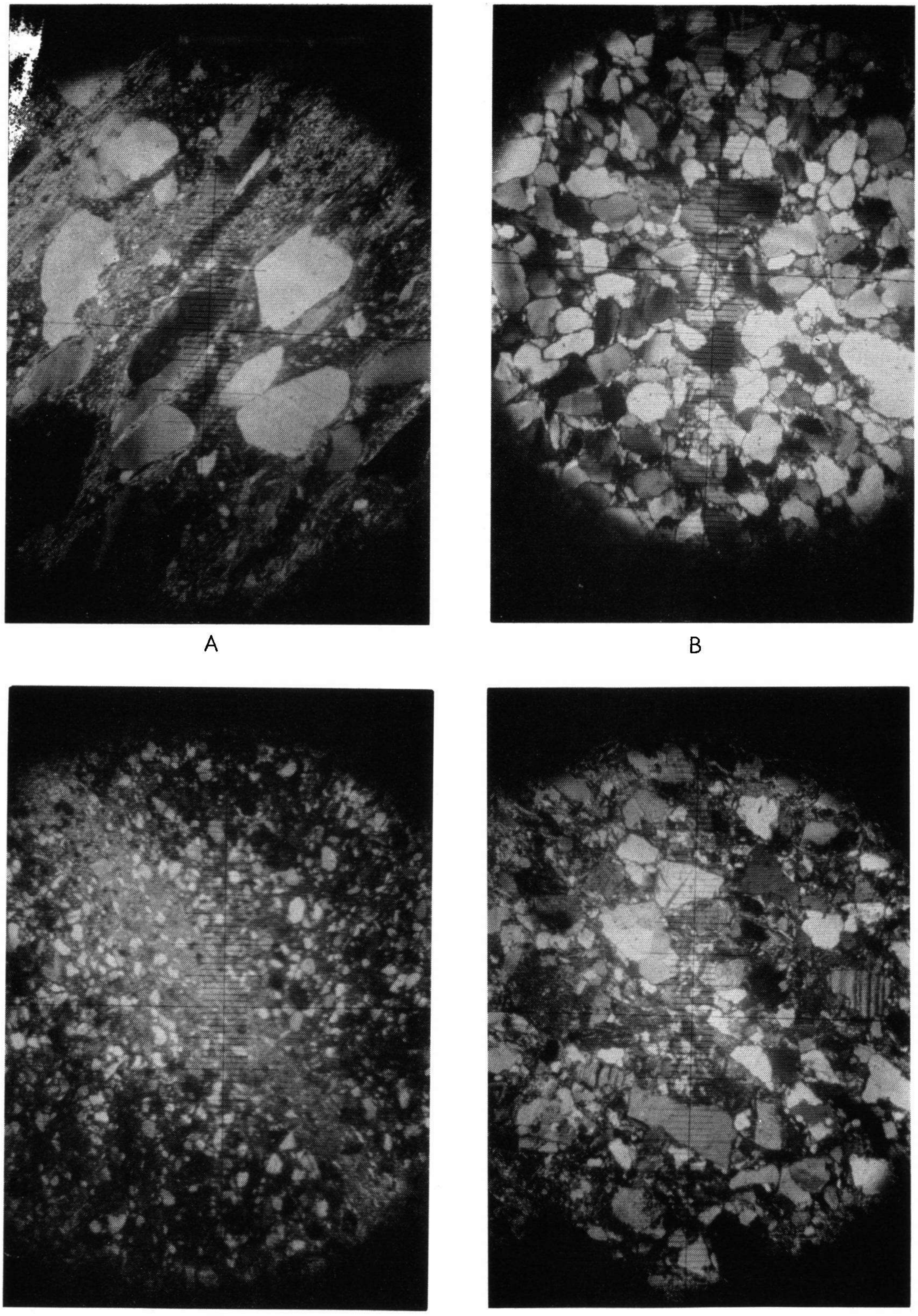


Fig. 58. Microscopic appearance of contrasting sandstone types (one division equals 0.1 mm)
A) Cambro-Ordovician protoquartzites (Puerto de la Bonaigua)
B) Devonian orthoquartzites (Corbison)
C) Devonian sandstones (Vilach)
D) Carboniferous greywackes (Salies)

of homogeneous sandstone beds, predominance of graded units and predominance of other sedimentary structures. Fig. 58-C is a photograph of a thin section which is representative of the sandstones of the turbidite association.

In the area of apparent predominance of homogeneous sandstones, shale is only subordinate. In the fine-grained sandstones lamination is occasionally well-developed. The thickness of the various beds (20—100 cm) is fairly persistent, as can be observed along the southern bank of the Rio Garona between Las Bordas and Viella.

Grading was only occasionally noticed in the field. Curving of cleavage planes is usually the best indication of graded bedding. However, in these

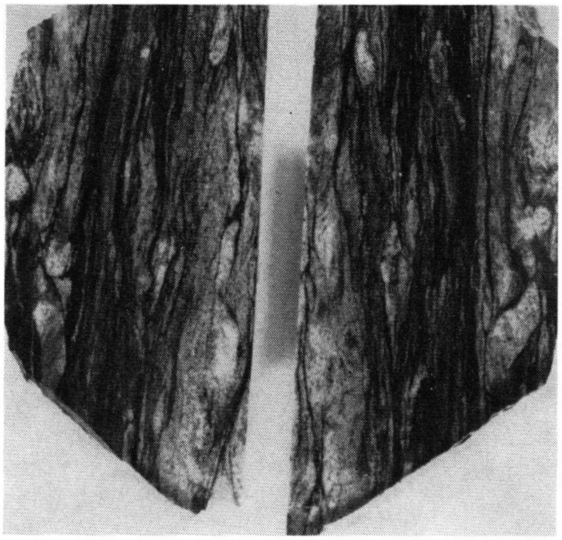


Fig. 59. Characteristic appearance of disturbed laminae in sandy beds, Corbison area

sandstones cleavage is poorly developed and therefore gives little assistance towards the recognition of grading.

North and north-east of the area of homogeneous sandstones, sandstones, siltstones and slates seem to be equally distributed. Most of the sedimentary structures are found in he central part of the turbidite association. Here the sandstone beds often consist of a succession of graded units.

East of the Rio Iñola most of the graded beds are graded siltstones, the grading being indicated by the slightly curved cleavage planes.

Neither grading nor other sedimentary structures have been observed in the eastern part of the turbidite association. South of Montgarri, a zone of homogeneous and poorly bedded sandstones occurs among siltstones and silty slates.

Limestones, matrix clay and grain percentages in thin sections of homogeneous sandstones of several regions have been compared; the average values of these percentages are shown in Table IV. The sub-angular grains are fairly regularly distributed in the matrix and consist of quartz with very little felspar. Variation in grain size is fairly slight. The medium

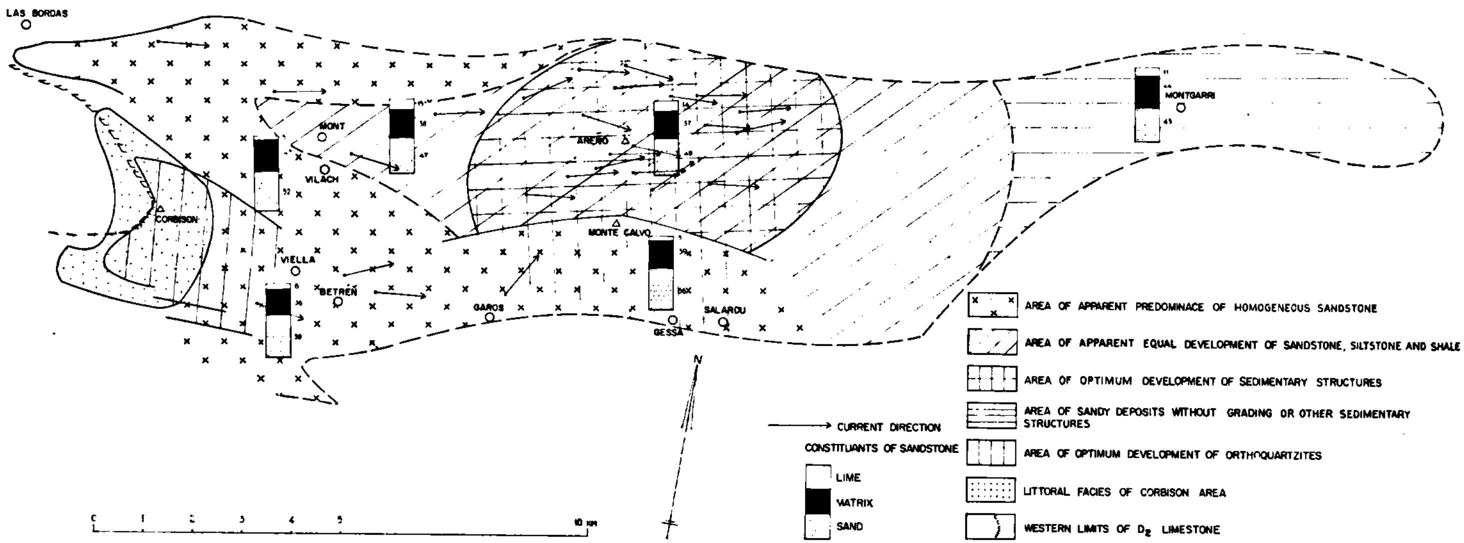


Fig. 60. Litho-facies map of Las Bordas formation

grain size of one graded sandstone bed is 35 micron, the coefficient of sorting 2.10, the coefficient of skewness 1.22 and the kurtosis 0.22. The average medium grain size of all the turbidite sandstone samples is 35 micron, the coefficient of sorting 1.54, the coefficient of skewness 1.12 and the kurtosis 0.31 (Pettijohn, 1957, p. 38). It should be noted, however, that these sorting coefficients are not comparable with data from sieve analyses as in this case only the coarser parts of sandstones or siltstones are included.

Patches of calcareous carbonaceous slates are sometimes found between the sandstones. In a locality north-east of Mont such slates are accompanied by very thin limestones; north of Aubert and west of Gausach carbonaceous patches are found in between homogeneous sandstones. Also south of the boundary between the Las Bordas and Viella formations, north of the Rio Garona (between Viella and Garos), patches of carbonaceous slates are found between light-coloured slates.

c. *Molasse association*

The Carboniferous is represented by a molasse association as conjectured by Pettijohn (1957, p. 618—622). The bulk consists of micro-conglomerates and medium to coarse grained ill-sorted greywacke beds showing angular grains. Fine-grained sandstones, sub-arkoses and shales are merely subordinate. The ill-sorting and the angularity of the grains are characteristic of this association, and well-rounded pebbles only occur in the few conglomerates present. Average compositions of this association are shown in Table IV and a photograph made from a thin section of a representative greywacke sample is shown in Fig. 58-D. Owing to the difficulty access to the deeply dissected terrain it was not possible to construct a continuous lithologic secion of the Carboniferous, although short sections were measured at two localities. The section in the Coll del Puis (Fig. 13) may be representative of the lower part of the molasse association; the other section along the Els Neres (Fig. 13) is located about 500 m above the base of the molasse association. Certain lithologic units can be identified; these units are indicated in the sections and estimated percentages of their constituents are given.

It should be noted that in general the greywacke beds are thicker when the grains are coarser. The coarse greywacke beds and micro-conglomerates do not form very persistent layers. Thin shale intercalations in the fine-grained greywackes often wedge out nor are these greywackes generally very persistent.

B. SEDIMENTARY STRUCTURES

a. *Graded bedding*

Table V shows the different sedimentary structures occurring in both the turbidite and molasse association.

The succession B, C and D of a single bed (Fig. 61) represents a graded unit of which only the bottom part B shows grading. A graded bottom part of this kind will be referred to as a graded sandstone bed.

Examination of thin sections made of several sandstone samples showed that much of the beds is actually sandy siltstones, siltstones, sandy slates and silty slates.

TABLE IV

Average composition of Cambro-Ordovician, Devonian and Carboniferous Sandy Deposits.

	Quartz (grains)	Matrix clay	Carbonate	Felspar (grains)	Rock fragments	Nu. of samples
CAMBRO-ORDOVICIAN						
Proto-quartzites						
Pique Valley	44 %	54 %	—	2 %	—	3
May de Bulard	64 %	36 %	—	—	—	2
Bonaigua	66 %	30 %	—	4 %	—	3
Marimaña	70 %	30 %	—	—	—	3
Micro-conglomerates						
Pique Valley	72 %	28 %	—	—	—	3
Bonaigua	64 %	36 %	—	—	—	4
DEVONIAN						
Ortho-quartzites						
(Las Bordas formation)						
Vilach	98 %	2 %	—	—	—	2
Corbison	100 %	—	—	—	—	2
Gessa	97 %	2 %	—	1 %	—	2
Salardu	98 %	—	—	2 %	—	2
Homogeneous sandstones						
(Las Bordas formation)						
Vilach	52 %	43 %	5 %	—	—	12
Rio Negro	57 %	36 %	6 %	1 %	—	8
Areño	56 %	39 %	5 %	—	—	12
Montgarri	45 %	44 %	11 %	—	—	8
Bottom part of						
Graded sandstones						
Vilach	47 %	38 %	15 %	—	—	28
Areño	49 %	37 %	14 %	—	—	20
Homogeneous						
Viella sandstones						
Rio Negro	40 %	59 %	—	1 %	—	4
CARBONIFEROUS						
Greywackes	39 %	31 %**)	—	17 %	13 %	6
Conglomeratic greywackes*)	38 %	39 %	—	14 %	9 %	4
Sandstones	24 %	76 %	—	—	—	2
Sub-arkoses	64 %	20 %	—	16 %	—	2

*) Sand-matrix only
**) detrital biotite 7 %, muscovite 2 %

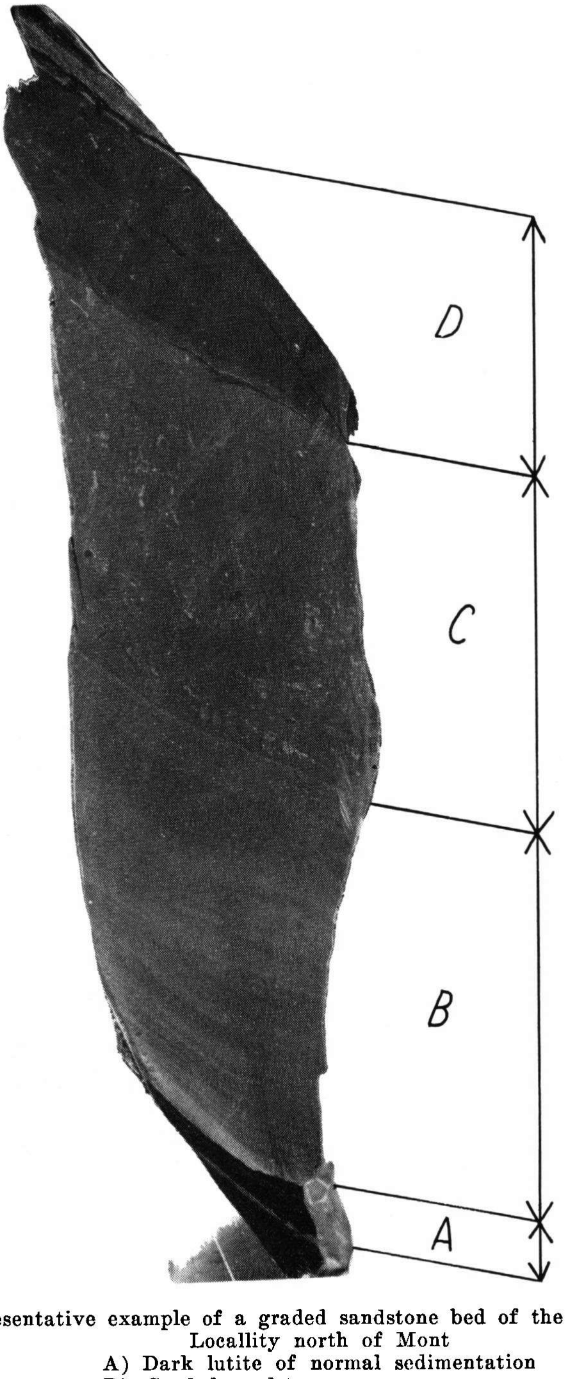


Fig. 61. Representative example of a graded sandstone bed of the turbidite association.
Locallity north of Mont
A) Dark lutite of normal sedimentation
B) Graded sandstone
C) Homogeneous silty slates
D) Slate

TABLE V

Occurrence of sedimentary structures in sandstone types of the Carboniferous molasse association and the Devonian turbidite association

	CARBONIFEROUS	DEVONIAN	
	Greywacke	Sandstone	Siltstone
Graded bedding	rare	common	common
Cross-lamination *)	—	observed	common
Convolute lamination	—	observed	common
Flute casts	—	observed	—
Drag marks	—	observed	—
Cross-lamination **)	common	—	—
Cross-bedding	common	rare	—
Ripple marks	common	—	—
Load casts	common	observed	—
Problematic current markings:	observed	observed	—

*) associated with turbidites
**) not associated with turbidites

Lamination is one of the main characteristics; generally present from bottom to top (Fig. 65-B), but in most of the graded units it seems to be better developed towards the top.

In the central part of the turbidite association graded units alternate

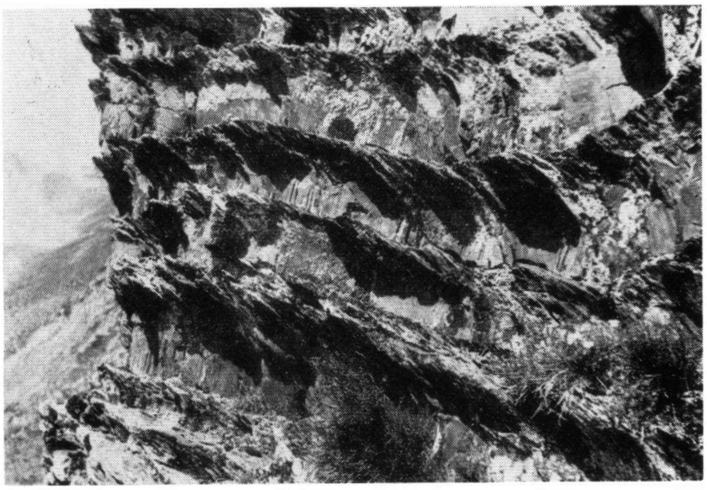


Fig. 62. Continuous graded beds, right side up, exposed near Areño

with homogeneous sandstones, siltstones and slates, but repetitions of graded units of varying thickness are frequent (Fig. 62).

Gravitational sorting, largely depending on weight, size and shape of the particles, should be expressed in grading. In the field, however, grading was only occasionally noted in the fine-grained sandstone beds as a gradual change in colour; the darker the colour the finer was the grain. But even the finest grading could be recognized by the curvature of cleavage planes.

For further information on the grading in fine-grained sandstones with only small grain size variations an examination was made of thin sections perpendicular to the bedding planes of several of these sandstones.

Statistical analyses of grain counts in thin sections showed that many apparently homogeneous and laminated sandstones are in fact fairly well-graded.

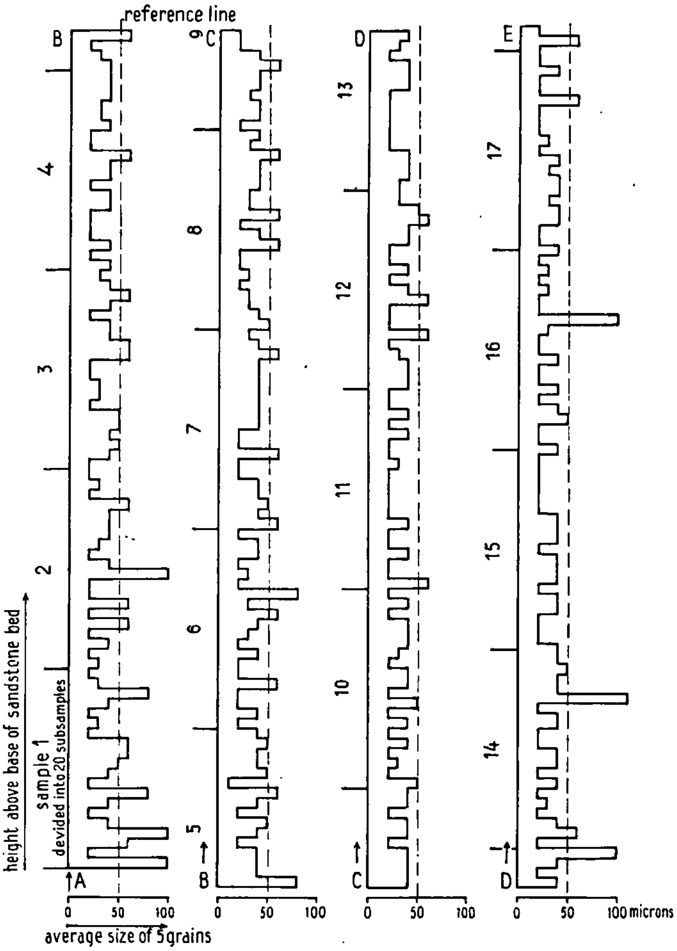


Fig. 63. Graph of grain size average from bottom to top of a very fine-grained, graded sandstone bed
The vertical distances are not to scale; equal distances are used for each sub-sample, but the height of a sub-sample may vary between 100 and 200 micron

The total thickness of the sandstone part of a graded bed was divided up into ten to twenty samples each consisting of 10 to 20 sub-samples of 5 grains measured. The average grain size of the 5 grains in such a sub-sample was plotted against height above base of a sandstone bed (Fig. 63). The number of sub-samples coarser than a certain reference size was counted in the successive samples. In most cases significantly negative Kendall's rank correlation coefficient indicated the relation between this measure of coarseness and the height of the sample in the sandstone part of a graded bed (Siegel, 1956, p. 213). This shows that there is a gradual decrease of average grain size towards the top of the bed.

The volume not occupied by quartz and felspar grains is filled by the matrix. The graded sandstones beds show wide variations in grain and matrix percentages (Fig. 64). The estimate of matrix content is based on area percentages, and as the thin sections cut at random through the grains

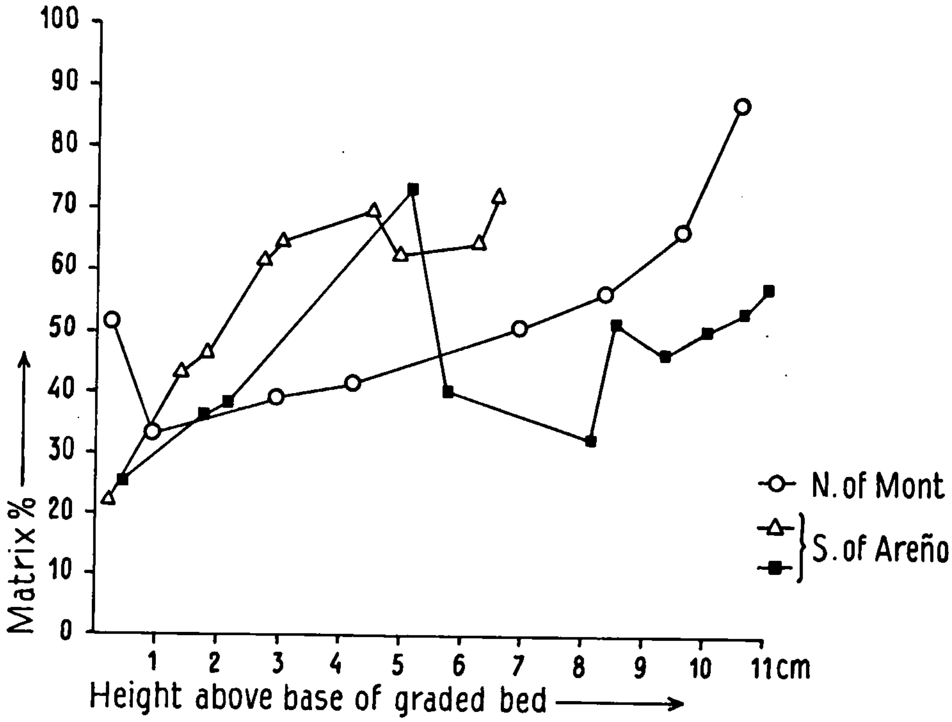


Fig. 64. Variation of estimated matrix contents within a vertical section through a graded sandstone bed

(dispersed at random in the matrix) such estimates of matrix percentage are fairly reliable (Chayes, 1954). In Fig. 64 a slightly laminated graded sandstone bed from a locality north-east of Mont shows a fairly regular increase of matrix percentage from bottom to top, except near the base. The irregular increase of matrix percentages of two highly laminated graded sandstone beds of a locality south of the Pico de Areño results from these laminae.

Fig. 61 is an example of a slightly laminated graded unit consisting of a graded bottom part (B), a homogeneous silty part (C) and slate (D). A thin layer of very dark-coloured lutite occurs overlying the slate part in sharp contact and underlying the graded bottom part in irregular contact (A). The different appearance and irregular contact of the lutite may indicate that it is not associated with the graded unit in between. These layers of dark lutite presumably represent the "normal sedimentation" repeatedly interrupted by the graded units resulting from turbidity currents.

The carbonaceous patches occasionally observed in the area of homogeneous sandstones differ in chemical composition from the dark lutite. A comparison of the chemical analyses (Table I) of such a carbonaceous patch (sample 2) with the shaly upper part of a graded unit (sample 3) containing dark lutite and a very dark-coloured Ordovician slate (sample 6) shows that the high carbon content of sample 2 does not agree with the very low carbon content of the "normal sedimentation" of lutite. Carbonaceous slates in the Pyrenees are only characteristic of the Silurian developed in a "black shale" facies and are also occasionally found in the Carboniferous. These carbonaceous patches are presumably also deposits, originating from eroded Silurian beds.

b. Cross-bedding and cross-lamination

Small-scale cross-bedding is frequent in the molasse association and is characteristic of the upper part of the Entecada formation in the Rio Iñola area.

Cross-lamination with a height up to 15 cm is occasionally found at the base of the turbidite association east of the Pico de Areño. In this area cross-lamination of the upper part of silty beds is common (Fig. 65-A).

At the top of some of the sandstone beds in the area of homogeneous sandstones a wavy lamination occurs filling shallow bowl-shaped depressions. These depressions succeed each other in the form of a festoon.

In the molasse association irregularly shaped fine-grained sandy laminae alternating with shale often show very delicate cross-lamination (Fig. 66).

c. Convolute lamination

The wavy appearance of persistent laminae known as convolute lamination is generally found in the upper part of sandstone beds in the turbidite association. Narrow "anticlines" varying in height from 2 to 20 cm are separated by wider troughs. Current direction can be inferred from overturned "anticlinal" crests consistently in one direction.

Although abundant in the Pico de Areño area (Fig. 67) convolute lamination also occurs west and south-west of the area occupied by the turbidite association. At several localities it was found possible to expose the convolution along the crest over a distance up to 15 cm. As a rule the crest-line appeared to be straight and was used for measuring the current direction.

d. Flute casts and drag marks

Regularly shaped, oblong flute casts were found on stray blocks. The casts are up to 10 cm in length. They should be classified as linguiform flute casts (Ten Haaf, 1959-b, p. 28). Much smaller markings, resembling flute casts, were found in situ. They were, however, difficult to distinguish from small-scale contortions resulting from severe tectonization.

Drag marks showing a relief of about 1 cm were found on stray blocks, and were accompanied by smaller and less regular drag marks running more or less parallel. Here confusion of weakly developed drag marks in situ with tectonic lineations is often unavoidable.

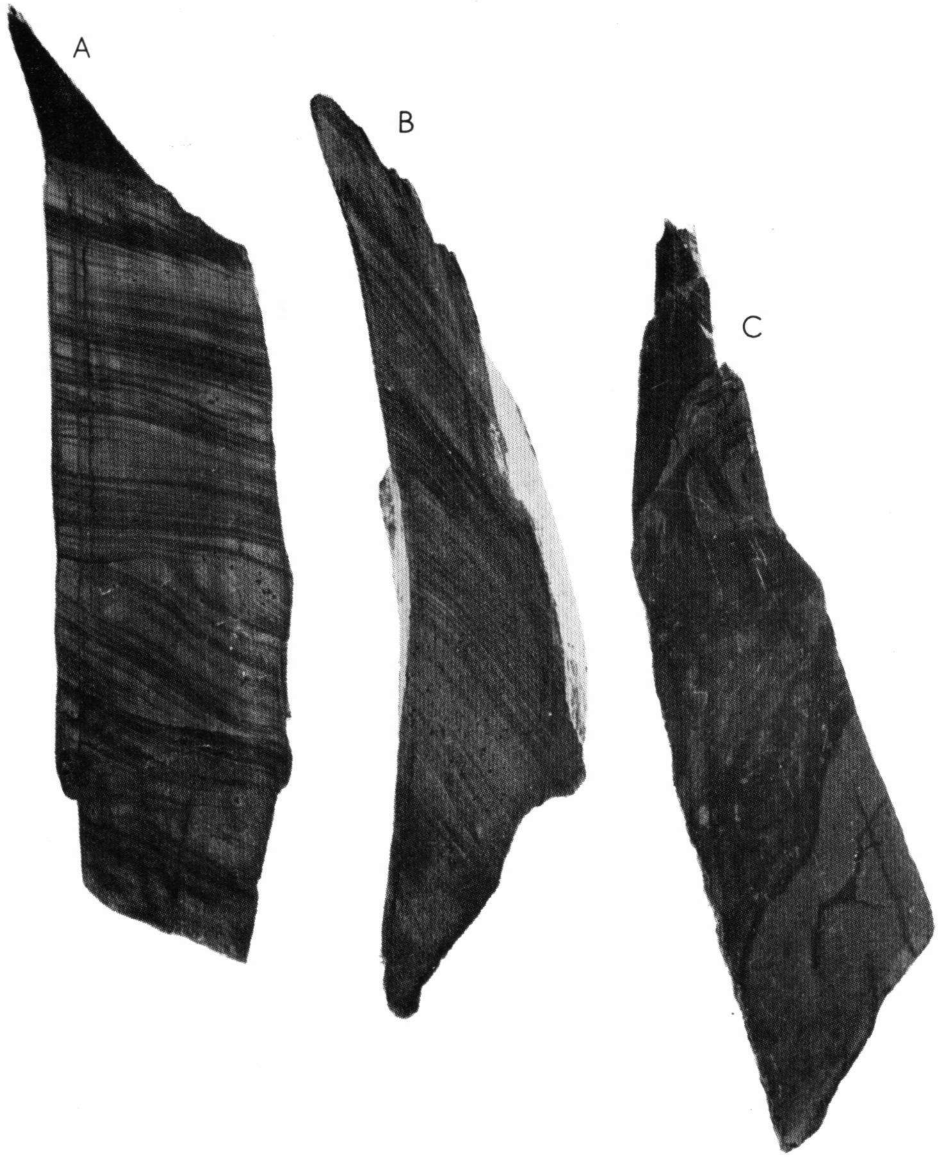


Fig. 65. Examples of various appearance of structures in sandstone beds of the turbidite association, Pico de Areño

A) bed showing well-developed cross-lamination
B) laminated bed
C) bed consisting in lower part of almost homogeneous sand, upper part completely contorted

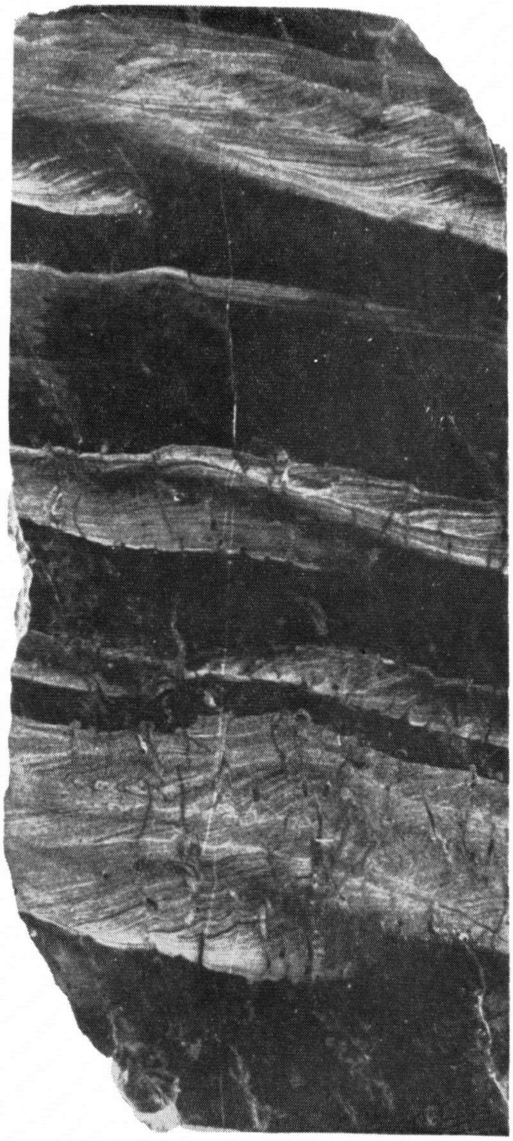


Fig. 66. Cross-lamination in irregular, thin sandstone beds alternating with shale. Note consistency of direction of dip of the laminae, very small-scale load casts at the sharply defined base of some of the laminated beds. In shale burrowings showing as sand patches. Carboniferous, Upper Rio Negro

e. Ripple marks

In the molasse association small-scale ripple marks are frequent exposed on the surface of shaly beds often over the entire length of an outcrop.
Tiny asymmetric current ripple marks occur which greatly resemble ripple marks such as are found on sandy beaches. Hardly recognisable markings in the turbidite association east of Las Bordas are presumably current ripple marks.

f. Load casts

Load casts, mostly small in size, frequently occur in the molasse association. Their length parallel to the bedding is usually less than 10 cm and may even reach micro-dimensions, as can be seen in Fig. 66. They appear as semi-circular and often as distinct saggings of coarse sand into mud. Groups of small-scale load casts occasionally give a strong impression of being oriented. Seen in the bedding plane these load casts are persistently asymmetric.

Fig. 67. Convolute lamination

Load casts are much less frequently observed in the turbidite association and are usually of a different nature. Although in general much larger they are usually less pronounced.

g. Problematic current markings

Between Las Bordas and Vilach current markings are found in a very regular arrangement (Fig. 68). They appear very much the same as symmetrical longitudinal ripple marks (Ten Haaf, 1959-b, p. 22). The oblong markings measure between 10 and 20 cm and show a relief of approximately 1 cm. Their longest axis is in complete accordance with the general current trend.

C. CURRENT DIRECTIONS

Measurements of current directions was hampered by marked folding, poor exposures and obliteration of sedimentary structures by cleavage. Most directions were obtained from cross-lamination measurements and from measurements of the crest-lines of overturned convolute lamination.

Use was made of Kopstein's method of measuring cross-lamination (1954, p. 47—56). It was not possible to measure cross-lamination traces on two faces of a bed transecting each other at right-angles. As the lamination traces hardly can be seen on fresh faces, the weathered natural faces afforded the best facilities for taking measurements. In order to avoid as far as

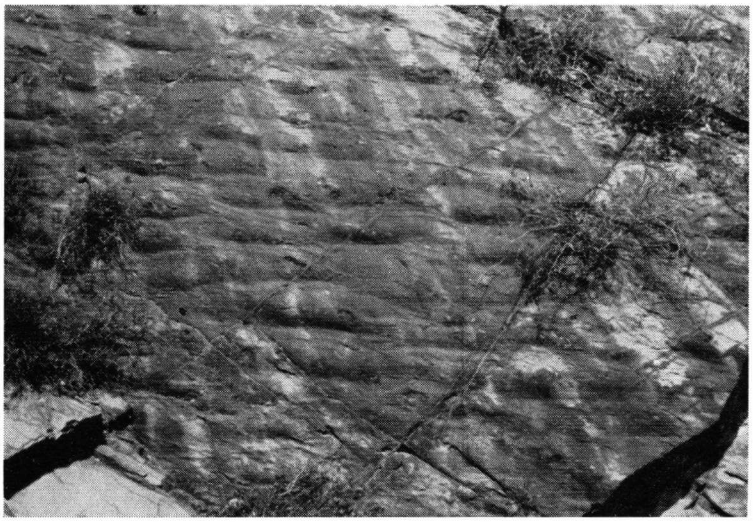


Fig. 68. Problematic current markings west of Aubert

possible measuring on faces belonging to one joint system differently oriented faces of one bed were measured. The small size of the folds necessitated making corresponding bedding measurements for every measurement or set of measurements of the cross-laminations.

Fold plunges could generally be measured with sufficient accuracy from lineations resulting from cleavage and bedding intersection. But variations in axial dips are frequent at fairly short intervals, so that a considerable error may still be introduced in the corrections needed for plunge. Dip and plunge corrections were applied by means of stereographic projection.

Average directions for each bed or for several beds together in large outcrops were obtained by vector summation. Only the central part of the turbidite association, the Areño area, yielded sufficient measurements for constructing significant direction vectors.

The measurements clearly indicate an eastwards directed transport of sediment (Fig. 60). The oblong shape even remains after reconstruction of the original basin which is shortened about 50 % in a north-south direction.

As the general current direction coincides with the longest dimension of this sediment accumulation it may be regarded as an example of longitudinal filling of a turbidite basin (Kuenen, 1957).

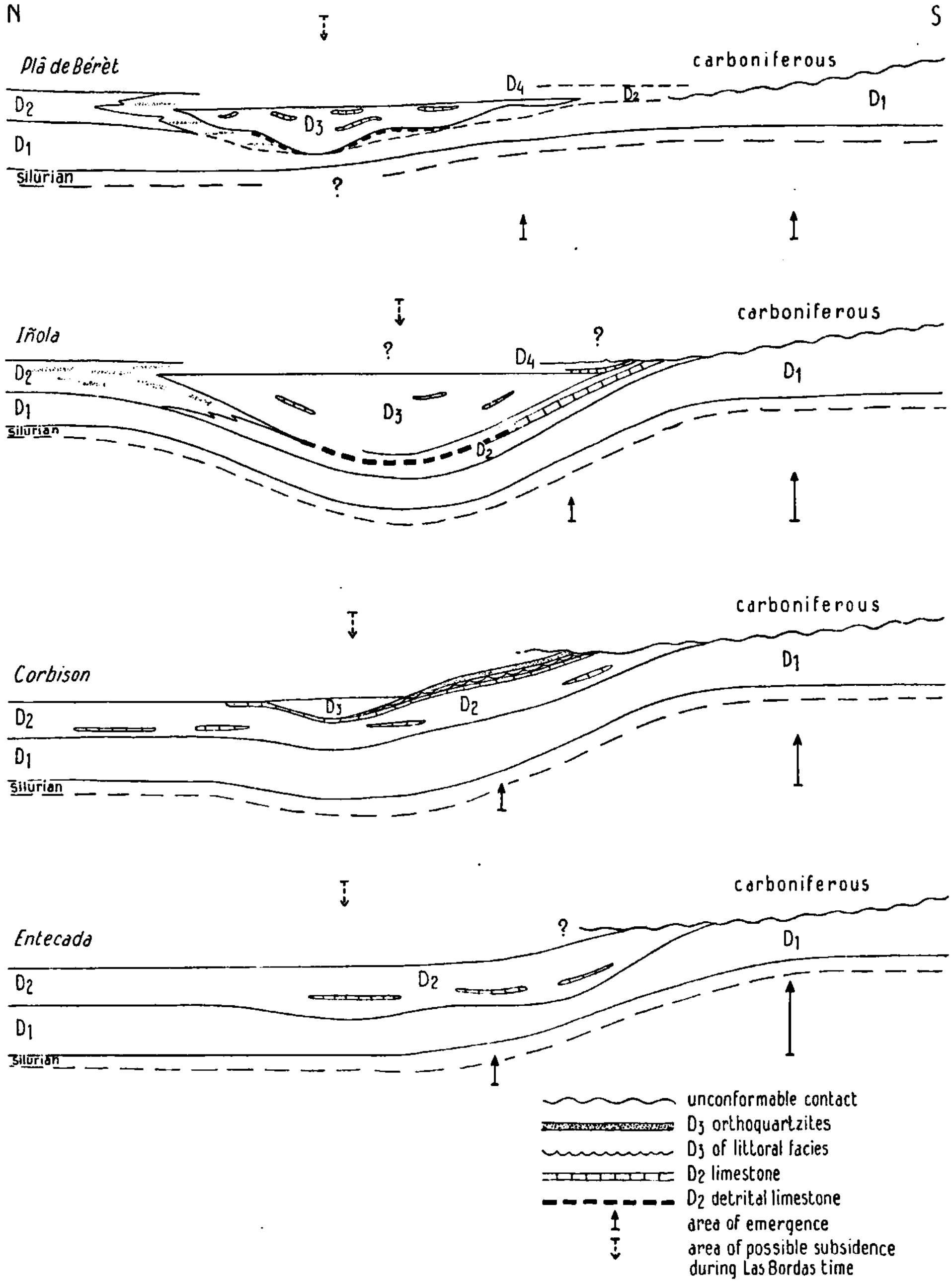


Fig. 69. Sedimentation history of the Devonian

Cross-bedding, oriented load casts and ripple marks were frequently measured in the molasse association. A large spread in current directions was observed, but there appear to be certain preferred directions to the north and east.

D. HISTORY

Generally speaking, sedimentation of the Cambro-Ordovician took place in a shallow marine to neritic environment. Table IV shows the composition of some Cambro-Ordovician sandstone samples. Alternation of irregular, thin layers of shale and sandstone or gravel is typical of the predominantly sandy sedimentation. Fig. 58-A, made from a thin section, shows a characteristic example of a Cambro-Ordovician sandstone.

Subsidence, usually keeping in step with rate of deposition, may have been occasionally interrupted. The resulting erosion has locally formed irregular layers or lenses of angular conglomerates (limestone conglomerates of the Pic de Past) or rounded conglomerates (Vallée du Port de Venasque). Situated between conglomerates the calcaire métallifère has been formed in a shallow marine environment. Crinoid remains, locally abundant in these metamorphic limestones, indicate an autochtonous and partly biostromal limestone, explaining the variations in thickness. Post-calcaire métallifère sedimentation becomes more pelitic in the south-west, south and east; in the north and north-east, however, the sand content is still predominant.

A very persistent facies type is characteristic of the Silurian. A very slow rate of sedimentation far from a source area of low relief may explain the very fine granularity and high content of organic matter of the black shales.

In the northern, eastern and southern part of the axial zone calcareous sediments constitute the majority of the Devonian. In the central part of the axial zone deposition in a deeper marine environment resulted in alternating shales and limestones (De Sitter, 1956-b). The history of Devonian sedimentation is outlined in Fig. 69. Differences in Devonian and Carboniferous sedimentation are shown in Table VI. The sea became shallower at the end of the Entecada times as is indicated by allochtonous current-transported sediments (slightly carbonaceous slates and cross-bedded limestones). Sandstone development in the Entecada formation in the north-east may indicate shallower conditions in this direction. South of the Las Bordas-Bagerque line a shallow marine orthoquartzite-limestone association developed thinning towards the east.

Emergence in the west and contemporaneous subsidence in the east resulted in down-sliding of accumulated sediments. Turbidity currents, repeatedly interrupting the normal sedimentation, filled longitudinally an oblong shaped trough (Kuenen, 1957, p. 191). Presumably the turbidite association derived its material from eroded sedimentary beds as is evidenced by patches of resedimented carbonaceous slates, absence of felspar and subangularity of the sand grains.

A period of erosion removed the Entecada and Las Bordas formation in an area south of the Central Anticline. Absence of this formation here may perhaps include non-deposition during the beginning of the Carboniferous in the central part of the axial zone. The Carboniferous of the Valle de Arán may therefore have a Namurian or lower Westphalian age.

In the northern and southern parts of the axial zone, however, a hiatus between Devonian and Carboniferous is only accidental and in general sedimentation was continuous (Ziegler, 1959).

Continuation of epirogenetic movements in the west and south-west during Carboniferous times resulted in rapid sedimentation of greywackes. Similar greywackes are found east of the Maladeta s.l. grading into marine slates towards the east.

A Triassic age has been assigned to the red-coloured sediments of the Pico de Salana, but there are too few outcrops to enable us to form a definite idea of the bulk composition of these sediments in this region.

The composition of the late-Miocene sediments of the small outcrops between the Aiguamoix and Arties valleys enables us to draw some conclusions regarding their formation and the relief of the source area. Erosion following uplift accounts for coarse sedimentation. The late-Miocene conglomerates, however, contain well-rounded pebbles, averaging 1 cm. The deposits otherwise consist of alternating sand, clay and lignite. Their composition favours a more extensive area of deposition under conditions capable of repeatedly forming very thick layers of peat. According to De Sitter (1956-b) sedimentation took place in a W.S.W.—E.N.E. river system. Indeed, subaquous deposition might partly account for the relatively fine sedimentation. But a more or less mature relief of the (nearby?) source area should be expected. Absence of granitic components in the conglomerates indicate deposition while part of the granodiorites still had a sedimentary cover and a fairly mature relief.

TABLE VI

Contrasting Devonian and Carboniferous sedimentation

Litho-stratigraphic group	Dominant lithology	Limestone development	Cyclical sedimentation	Fossil content	Sandstone type	Depositional environment (Lithofacies)
Carboniferous (Molasse association)	Greywackes	Thin limestone beds max. 10 m (total)	—	Plant remains	Fine to coarse-grained conglomeratic greywackes	Paralic
Viella slates and sandstones (D_4)	Shale	One known limestone bed max. 2 m	—	Not observed	Fine-grained sandstones	Shallow marine
Las Bordas Sandstones (D_3) (Turbidite association)	Sandstone/ siltstone	Detrital limestone and thin nodular limestones	Well-developed	Not observed	Fine-grained sandstones and siltstones	Deeper marine
* Las Bordas orthoquartzites (D_3)	Sandstone	No limestone	—	Not observed	Medium to fine-grained orthoquartzites	Very shallow marine
* Las Bordas "littoral facies" (D_3)	Sandstone/ shale	No limestone	—	Not observed	Medium to fine-grained sandstones	Very shallow marine
* Entecada limestone (D_2)	Limestone	Optimum	—	Rare crinoid remains	—	Very shallow marine
Entecada slates (D_2b)	Shale	Poor	—	Rare	Fine-grained sandstones	Shallow marine
Entecada slates (D_2a)	Shale/ limestone	Detrital limestone	—	Not observed	—	Very shallow marine
Entecada slates (D_2)	Shale	Occasional thin limestone beds	—	Rare	Fine-grained sandstone intercalations	Deeper marine
Basal limestone (D_1)	Limestone	Optimum	—	Rare crinoid remains	Fine-grained sandstone intercalations	Deeper marine

* Orthoquartzite-limestone association

CHAPTER IV

MORPHOLOGY

A. PLANATION LEVELS

High plateaus found in the Pyrenees have long been considered as well-preserved remnants of old denudation surfaces.

Platforms have been described and dated by many authors, but there is little uniformity either in dating or in correlation. Some authors even regard a plane which can be constructed over a number of summits of about equal height or "Gipfelflur" as evidence of an old denudation surface. Faucher (1938) supposes that this "Gipfelflur" in the Pyrenees is of Hercynian age; Birot (1935) attributes to it a pre-Alpine origin, whilst Garcia-Sainz (1940) attempts to define its age more precisely as Cretaceous.

In our area remnants of denudation levels, usually scarcely recognizable, are indicated in Fig. 70.

Planation surfaces are found roughly at three levels: 2,400—2,600 m, 2,000—2,200 m, the most extensive level, and 1,700—1,800 m. In the Sierra Negra a level at an altitude of 2,500 to 2,600 m is preserved, exclusively on a Silurian subsoil. Broad undulating ridges, strongly weathered, alternate with slopes which dip gently towards a 2,000 to 2,200 m level, the latter deeply dissected by river erosion.

Other platforms at similar heights are found in the extensive, hilly regions of La Montañeta and El Pruedo, which are partly covered with thick moraine deposits (Fig. 71). Flat valley bottoms are found between 2,000—2,200 m in the upper reaches of the Rio Iñola and Rio Esera.

A lower situated surface (1,700 to 1,800 m) is formed by the plateaus of Campsaur (463/48) and Superbagnères (456/52) which were recognized as such by Goron (1942), and in the Plâ de Bérèt. The plateau of Campsaur (Fig. 72) slopes gradually upwards to a ridge of 2,000—2,200 m, presumably the remnant of a planation level. The flattened tops of the Tuc de Media (2.200 m) (4°.31′/42°.41′) and the Pico de Baguera (2.400 m) (4°.40′/42°.42′) may also be remnants of erosion levels.

Late-Miocene deposits are found at an altitude of 1,820—1,980 m just below the platforms of El Pruedo and La Montañeta. Interpretation of a relation between planation surface and late-Miocene deposits is very difficult and open for two hypothetic possibilities of dating of the main planation surface.

According to De Sitter (1956-b) these deposits fill up a deeply incised gully in the main planation surface which therefore should be younger or of about the same age.

On the other hand the composition of the sediments is not in accordance with the sediments which could be expected in such a deep and narrow gully. The presence of longitudinal faulting (and the possibility of activation of such faults during repeated periods of elevation following the Alpine orogene) may indicate preservation in a faulted zone. This may lead to the assumption

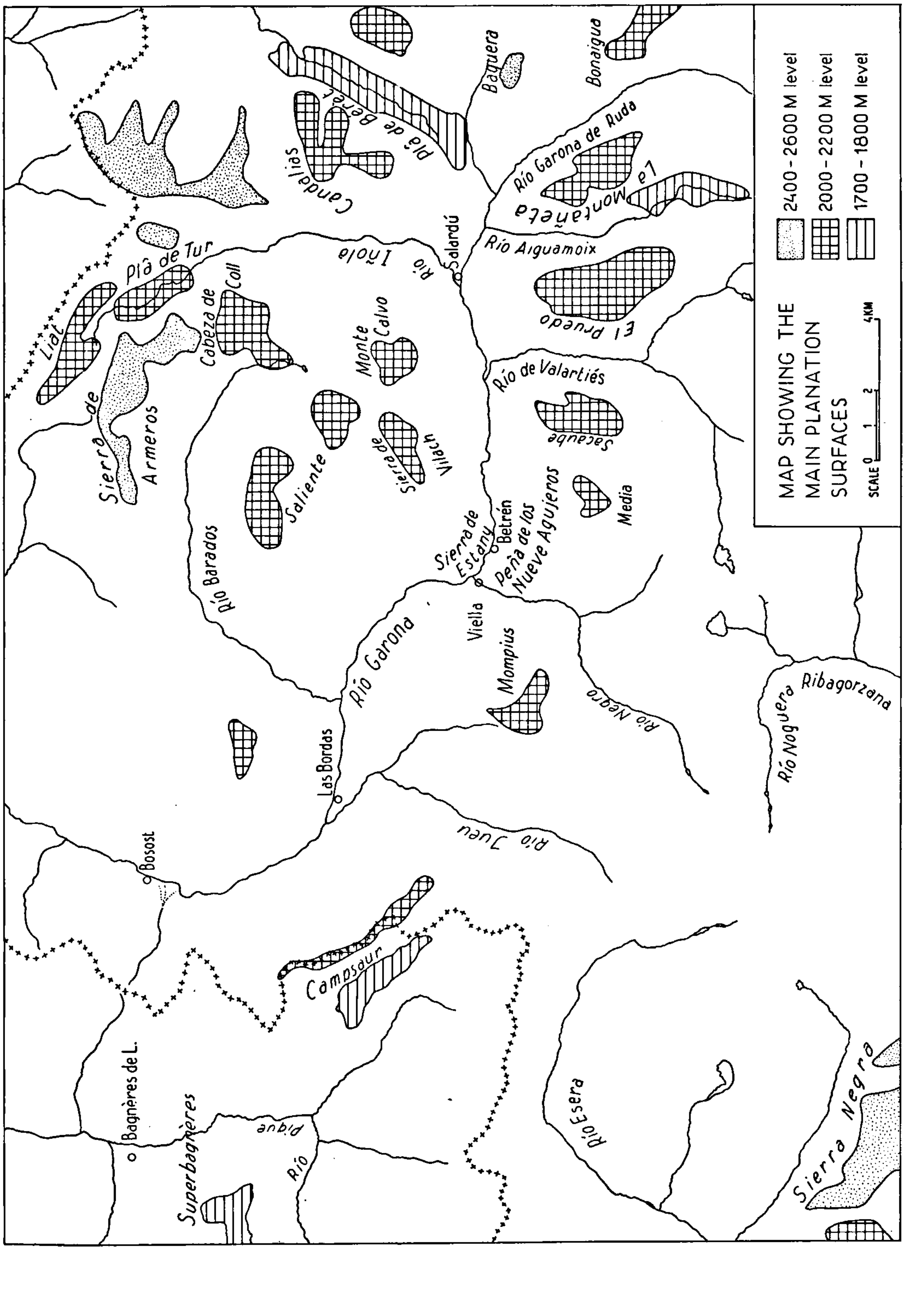

MAP SHOWING THE MAIN PLANATION SURFACES
2400 - 2600 M level
2000 - 2200 M level
1700 - 1800 M level
SCALE 0 1 2 4KM
Baquera
Bonaigua
Plá de Beret
Candelaliás
Río Garona de Ruda
La Montañeta
Salardú
Río Aiguamoix
Río Iñola
El Pruedo
Plá de Tur
Lial
Cabeza de Coll
Monte Calvo
Río de Valartiés
Sacaube
Sierra de Armeros
Saliente
Sierra de Vilach
Media
Sierra de Estany
Betrén
Peña de los Nueve Agujeros
Río Barados
Viella
Mompius
Río Garona
Río Negro
Río Noguera Ribagorzana
Las Bordas
Río Jueu
Bosost
Campsaur
Bagnères de L.
Superbagnères
Río Pique
Río Esera
Sierra Negra

Fig. 70.

of deposition of the late-Miocene in a larger area preceding or contemporaneous with the activation of faults. The extension of the main planation surface north and south of the flanking faults should then result from a subsequent period of erosion, indicating a post-late-Miocene (Pliocene?) age of this surface. Indications of faults flanking the late-Miocene deposits to the north, however, have not been found.

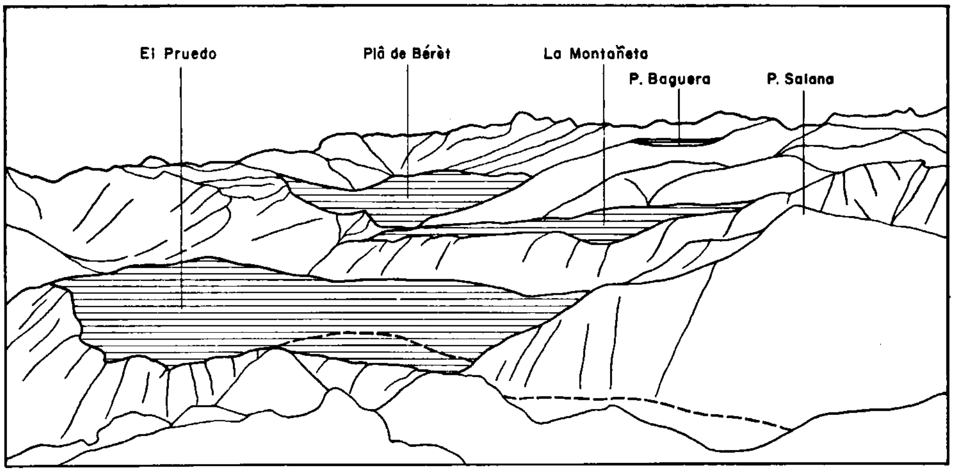


Fig. 71. The planation surfaces of El Pruedo and La Montañeta (2,000—2,200 m), Plâ de Bérèt (1,800 m) and the Pico de Baguera (2,400 m). After photograph

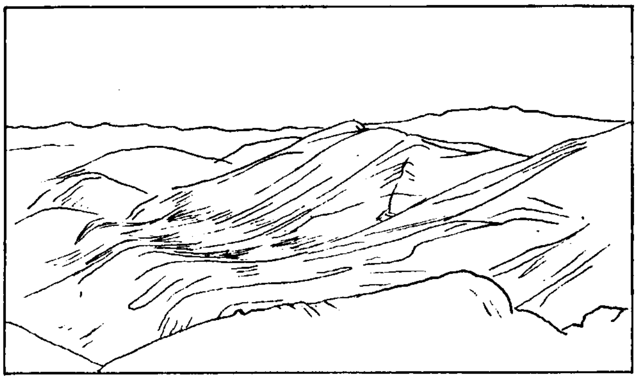


Fig. 72. Planation surface of Campsaur, seen from the Collado de Mounjora. After photograph

The strong uplift of the Pyrenees at the end of the Tertiary is followed by smaller movements during the Upper-Pliocene and the beginning of the Quaternary (Garcia Sainz, 1939).

Flat valley bottoms in the Rio Esera, Rio Aiguamoix and Rio Iñola found at similar altitudes (1,450—1,560 m) and probably other flat valley bottoms at lower altitudes might reflect these small movements.

B. GLACIATIONS

During he Quaternary period glaciers were formed and moved downwards, directed by the pre-existing valleys. The eroding influence of these glaciers, although much less than in the case of the Alpine glaciers, can still be seen in the many traces which are found almost everywhere. The convex form of the north slope of the Maladeta s.s. indicates a certain maturity of the pre-glacial topography, which was only slightly affected by the subsequent glaciations.

According to Nussbaum (1938) indications of at least two glacial periods are found in the northern Pyrenean foothills. Llopez Llado (1946) and Garcia Sainz (1941) distinguish three glaciations, of which the first may have been of the Scandinavian type leaving few traces. The two subsequent glaciations were of the Alpine type, the first being greater in extension.

It seems probable that in the area described the many glacial traces belong exclusively to the last glaciation. Moraines, occasionally found, probably represent only intermediate stages.

C. CLACIAL REMNANTS

Glacial striae, "roches moutonnées", rock steps and glacial lakes are frequent in the higher parts of the granodiorites, and somewhat less frequent in sedimentary areas where slates and sandstones predominate.

In the upper valley of the river Pique (463/45) different phases of the last glaciation might be deduced from three successive rock steps preserved in the east side of the valley.

Many glacial lakes are found in cirques, especially in the granodiorites, and generally reach great depth. Glaciers have formed several large but shallow mountain-pass lakes like Lago Rius (4°.29′/42°.38′), Lago Liat (481/56) (Fig. 73) and the smaller Lago de Toro (4°.22′/42°.40′). The small but deep Lago Barancs (4°.22′/42°.38′) in the Maladeta s.s. was formed by glacial erosion.

Aerial photographs clearly indicate how strongly the location and the shape, especially of the shallow and small lakes, are determined by joint systems. "Roches moutonnées" are also frequently found between intersecting joint systems.

The highest level of the glacier ice in various valleys could still be distinguished in the field by differences in weathering which aerial photographs show as a slight colour difference. By this method it was possible to estimate the thickness of the ice (Table VII).

Occasionally glaciers were connected to each other over the present watersheds. This has been proved for several glaciers (Table VIII). The Aiguamoix glacier derived its ice from El Colomes (4°.37′/42°.36′), the greatest glacial basin in the Pyrenees. This glacier was split near the Baños de Tredós by an obstructing peak which still rises 68 m in the middle of the valley. The ice stream partly covered the platform of El Pruedo and flowed to the Valarties. The eastern branch of the Aiguamoix glacier also covered part of the platform of La Montañeta. Both pre-glacial surfaces have a thick cover of moraine material.

The valleys of the upper reaches of the Garona (Iñola, Malo, Aiguamoix and Bargadera), except the Valarties valley, provide good examples of

TABLE VII

Greatest thickness of glacier ice

	Occurrence	Altitude	Maximum thickness
A.	Garona de Ruda	1,750 m	240 m
	Aiguamoix	1,850 m	220 m
	Aiguamoix	1,500 m	180 m
	Valarties	1,400 m	260 m
	Bargadera	1,800 m	220 m
	Esera	2,200 m	300 m
	Collado de Toro	2,200 m	150 m
B.	Garona at Viella	1,000 m	500 m
	Garona at Las Bordas	800 m	600 m
	Garona at Lès	700 m	700 m
	Pique at Luchon	625 m	600 m

A. Estimated with the help of aerial photographs.
B. Estimated from the occurrence of erratics.

hanging valleys. The Valarties glacier must therefore have been of about the same size as the combined Garona glaciers. Its large size may be due to the addition of part of the Aiguamoix glacier.

Glacial traces are found only sporadically in the valleys of the lower reaches of the Garona. East of Viella an unaffected remnant of glacial erosion is found in the rounded and convex Sierra de Estany north of Betrén (475/45). Its counterpart can be found south of this village in the hollow and concave form of the Peña de los Nueve Agujeros formed by the outer bend of the glacier (Fig. 70).

Granite boulders, sometimes very large, are frequently found on the valley slopes of the inner bend of the glacier but only sporadically on the valley slopes of the outer bend. These erratics may indicate the thickness of the glacier ice (Table VII).

TABLE VIII

Connected glaciers

Glaciers	Area of connection	Indications
Iñola and Pallaresa	Lago Montuliu area (485/54)	Glacial striae and orientated drumlins
Upper Esera and Garona	Collado de Toro	Glacial striae and U-shape of the Collado de Toro
Aiguamoix and Valarties	Plateau of El Pruedo	Moraine material
Garona and Pique	Colle de Portillon	Glacial striae

In the western valley of the Plâ de Artiga de Lin (4°.24′/42°.41′) we find among many glacial traces a well-preserved terminal moraine marking a regression phase of the last glaciation. The landscape north of the Plâ de Artiga de Lin is completely different; no signs of former glaciations are present except erratics of granite and micaceous sandstone.

Strong resistance to glacial erosion is offered by the Devonian sand-

Fig. 73. The Liat Lake and the Forat de Liat situated on the main planation surface. The pronounced line between lake and mountain top indicates the Liat fault

stones and quartzites near Viella. Granites and pegmatites also strongly resisted glacial erosion. In these rocks the valleys are narrow but broaden considerably in the soft mica schists near Bosost, Lès and Bagnères de Luchon.

Capture of the Rio Malo, originally a branch of the Rio Noguera Pallaresa, probably took place during the glacial period. The Pallaresa glacier obstructed the ice cover of the Plâ de Bérèt which in turn obstructed the Malo glacier, forcing it to flow south-west to the Garona. This is indicated by glacial striae. The Plâ de Bérèt, with its broad and slowly declining pre-glacial surface, is hardly affected by glaciation. Erratics are occasionally found.

The last phases of the retreating glaciers are found in periglacial features. Semi-circular moraines (epiglaciares) composed of loose blocks, are situated in the higher parts of the mountains in the debris of preceding glaciations. In the Sierra Negra sickle-shaped walls composed of Silurian debris are found at an altitude of about 2,100 m (Fig. 74). At a first glance they look very much like barchanes; presumably they were formed by the melting glaciers.

D. POST-GLACIAL EROSION

Large, rapidly-flowing water masses formed during the melting of the ice quickly removed the thin cover of moraine material left by the glaciers. This is shown in the lower reaches of the river Aiguamoix where the headward erosion reached the solid granite at a depth of about 10 m. In the valley of the Garona de Ruda and on the Plâ de Bérèt aerial photographs indicate clearly the points reached by the headward erosion. Moraine material is often found downstream of flat valley bottoms. Here the water cut deeply into the country rock, modifying the former U-shape of the valley into a sharp V-form. Mechanical weathering played a prominent role in the disintegration of the rocks. Vast cones of debris originated as soon as the pressure of the melting ice upon the walls of the glacial trough ended and the rock became exposed to atmospheric disintegration. A good example is provided by the upper reaches of the Rio Negro and Rio Valarties. Vast cones of dry debris fill up part of the valley, marking the retreat of the glaciers. Spreading over the valley bottom this debris changed the original U-shape into a

Fig. 74. Sickle-shaped periglacial remnants in the Sierra Negra. The area is covered with fine Silurian debris

semi-circular one (Fig. 75). Small, steeply-falling torrents transported great amounts of material into the main valleys, forming fine examples of alluvial fans in which the main stream later incised its course.

The inability of the Rio Malo to cut deeply into the slates before reaching the Garona may also account for the previously mentioned capture of this river during the preceding glaciation.

Lakes formed by the melting ice in the flat valley bottoms (deepened during the preceding glaciation) are quickly filled with fluvio-glacial material. Thresholds downstream of these plains are cut to a certain depth determining the river level upstream. Here the rivers have the wide, braided courses observed in the rivers Iñola, Aiguamoix, Esera, Ribagorzana (4°.27′/42°.38′) and Tort (4°.33′/42°36′). Many flat valley bottoms and the highest part of the Plâ de Bérèt are very marshy; locally thin layers of peat are exposed along the rivers.

Water disappears in sink-holes in the upper course of the Rio Iñola and the Rio Esera, but when the melting snow supplies large amounts of water the drainage capacity of the sink-holes is sometimes too small and the rivers follow the valleys.

An unique example of the eroding power of water is given by the

Guëlls de Jueu (467/43). Upstream of this enormous spring the glacial landscape has hardly been changed. Below the Guëlls the amount of water suddenly increases considerably, sufficient to cut a deep, narrow gorge, and to eradicate all signs of the preceding glaciation.

Terraces are found in the rivers Ruda and Aiguamoix at about 2 m above the river level. The Rio Iñola in its lower course cut out several terraces in extensive alluvial fans. Along the river Garona, between Viella and Salardu, terraces are situated 20 to 40 m above the river.

Fig. 75. The semi-circular shape of the Aiguamoix valley

Solid, thick-layered sub-recent slope-breccias cemented by limonite, are found locally in the Silurian of the Cabeza de Coll and the Sierra Negra. The long, gently-dipping slopes of the Sierra Negra are entirely covered with fine black flakes over which rainwater or meltwater runs down in broad fans forming such breccias. A slope-breccia of well-layered limestone components is found at an altitude of 1,800 m along the Baranco de Comasera (4°.25′/42°.42′), an eastern affluent of the Rio Jueu.

In the granodiorites and hard Carboniferous sandstones joint systems are responsible for the development of vast fields of dry debris. This is due to the loosening of great blocks by freezing and thawing. The arêtes above the cirques are therefore often sharply toothed.

But the relief adjusts itself to the lithology in even greater detail. Each different kind of rock shows its own characteristics of weathering. The more resistant rocks are the granodiorites and the quartzites sticking out as ledges on the east slope of the Corbison. The less resistant rocks are the soft Silurian slates, which generally form the depressions, as illustrated in the Silurian flanking the Central Anticline. The more resistant rocks include limestones; they are somewhat slowly dissolved, often showing a great variety of karst phenomena, especially the metamorphic limestone.

Talus cones on the steep east slope of the Rio Iñola extend from the top of the ridge to the river. On the ridge, and parallel to it, long and deep clefts are found in the slates and sandstones, marking future landslides. Occasionally, landslides bring down bedrock and large parts of the alluvial fans.

Surface creep occurs everywhere in the schistose rocks, especially on steep slopes where slates are present.

In the Cambro-Ordovician slates north of the Rio Malo, glacial striae have undergone vertical movements along joint planes. Vertical shifts of 1 to 3½ cm which form steps have been measured. Such small shifts would have been obliterated by glacial erosion had they existed before, and consequently the movements must have originated in post-glacial times.

E. HYDRO-GEOLOGY

Karst phenomena are typical of limestone formations, which in our region occur frequently in both the Ordovician calcaire métallifère and the Devonian basal limestone. These phenomena are especially developed in metamorphic limestones recrystallized in irregularly-shaped calcite crystals up to 5 mm in size, giving it a sugarly aspect. These marbles are easily dissolved by water. It has been noted, that sink-holes often occur at the boundaries between limestone and other formations, or that they are situated along faults.

Precipitation on limestone areas in the form of rain or snow disappears into sink-holes, and therefore small torrents, or lakes are rare. Torrents entering a calcareous area from elsewhere, quickly disappear, eventually finding a subterranean outlet.

Large areas are drained in this way; a good example is found in the upper-Esera region, where the meltwater from the glaciers of the Maladeta s.s. finds a subterranean outlet to the river Garona. The water disappears in the Forat *) de Aigualluts (462/42°.40′) (at 2,100 m) and in the Forat de la Renclusa (464/42°.40′) (at 2,225 m), then crosses underground the watershed between the rivers Esera and Garona formed by a 2,700 m high mountain range, reappearing at the other side in the Güells de Jueu **), situated on the Plâ de Artiga de Lin at 1,400 m (Fig. 76).

Penck in 1883 suggested a connection between the Forat de Aigualluts and the Güells de Jueu, in accordance with the general opinion of the local population. In 1931 this opinion was proved correct. In that year Casteret demonstrated the connection with the aid of fluoresceine. Later on (1951) De Lizaur y Roldan showed by the same method the connection of the Forat de la Renclusa with the Güells. Further investigations in this area gathered by our group has been published by B. G. Escher (1953) on the occasion of the first international speleological congress in Paris.

From a geographical point of view the subterranean capture of the upper Rio Esera is quite a noteworthy phenomenon. Its water reappears

*) = sink-hole.
**) = local name for a very strong spring meaning "Jupiters eye".

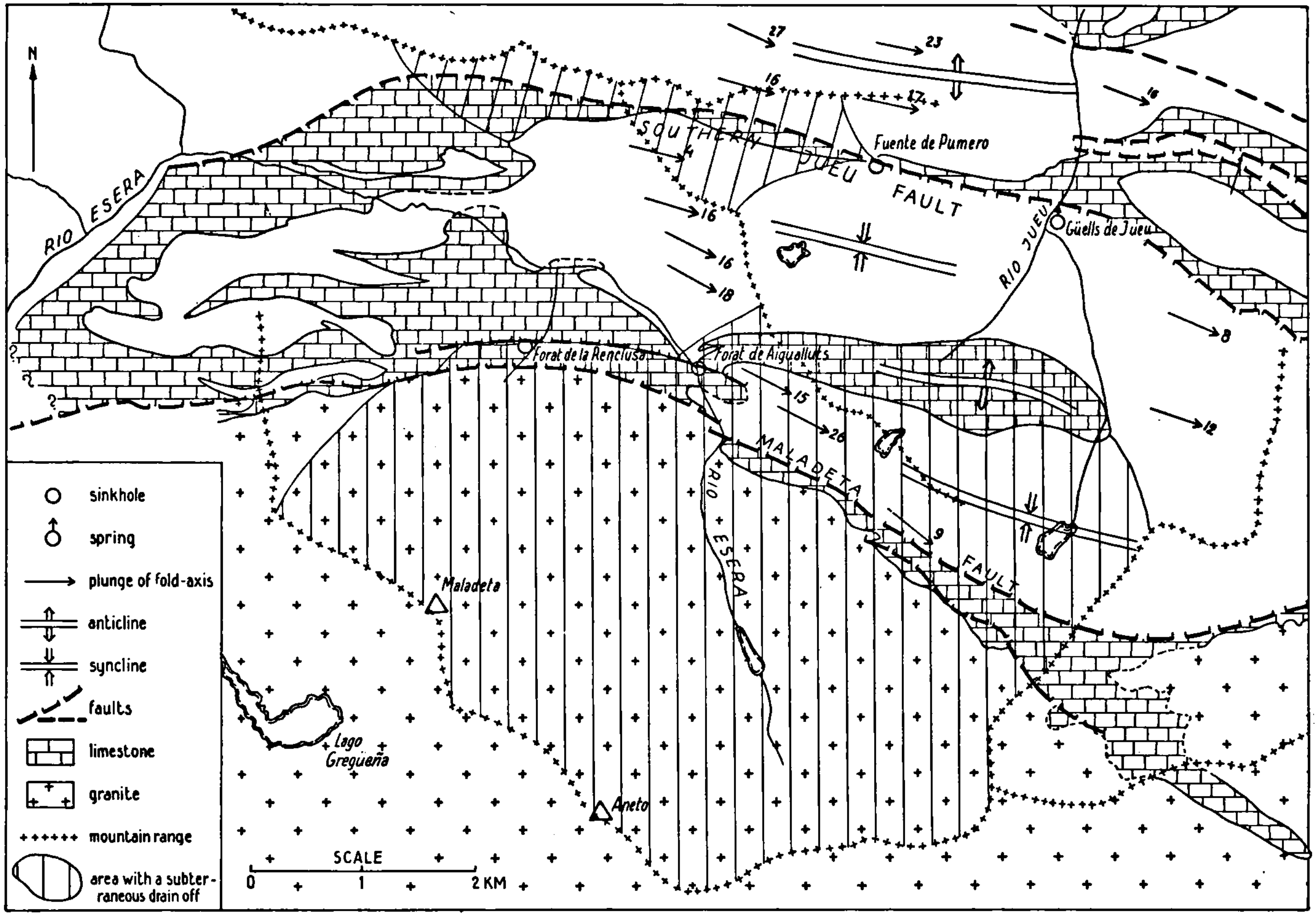


Fig. 76. Hydro-geological map of the drainage areas of the upper Esera and the upper Jueu rivers

at the other side of the watershed in another river, and thus water which normally would have flown to the Mediterranean is diverted to the Atlantic.

Geologically the connection is quite understandable. Both sink-holes in the Devonian basal limestone are situated on the same south flank of a

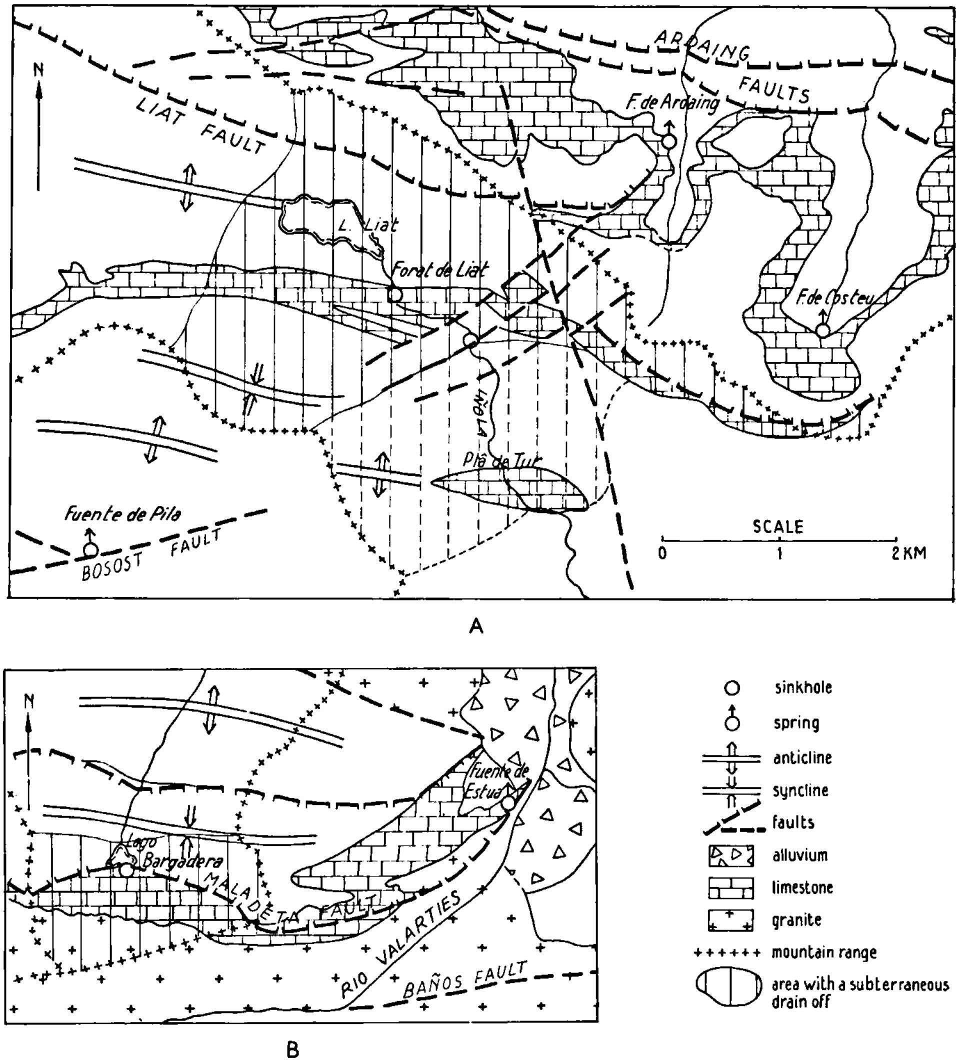


Fig. 77.

A) Hydro-geological map of the drainage area of the upper Iñola river and adjacent regions

B) Hydro-geological map of the drainage area of the upper Bargadera river and parts of the Valerties river

syncline, their position being determined by a fault (fig. 19, section 15 & 16). The Basal limestone and the overlying Carboniferous sediments form a syncline which plunges gently towards the east, and water, flowing along the bottom of the limestone or through it, is forced to flow eastwards.

The Güells (1,400 m) are situated in the middle of a moraine of large granite boulders (fig. 19, section 15); the water-bearing Devonian basal limestone was not found exposed here but crops out further to the north, where the southern Jueu fault separates it from the Carboniferous. In the Güells locality the water-bearing limestone forms a very probable culmination which has been partly eroded during the glacial period.

Another example is the Fuente *) de Pila (478/53) (Fig. 77-A), situated in the northern branch of the upper Rio Barados at 1,780 m in an outcrop of the calcaire métallifère. Some of the water is derived from a torrent flowing from Lake Liat, which disappears in an enormous sink-hole, the Forat de Liat (481/56), at 2,100 m. This sink-hole (Fig. 73) is situated between the limestone and the underlying sandstones. Downstream of the Forat de Liat more sink-holes occur along a fault line at 2,080 m which is probably connected below the surface to the same limestone, which is not exposed here.

It is not certain whether water also disappears at about 2,050 m in the calcaire métallifère in the course of the Rio Iñola south of the Plâ de Tur. Impurity of the limestone might have prevented the formation of sink-holes.

The Forat de Liat is situated on the steeply-dipping north flank of a syncline which passes the watershed between the rivers Iñola and Barados. Further south the structure changes into an anticlinal one, the water following the contours of this eastward-plunging anticline. Two faults, the Bosost fault and the Pila fault, cut this structure. The Fuente de Pila is situated at the junction of the two faults (fig. 18, sections 4 & 5).

North of the frontier range many sink-holes are also found in the calcaire métallifère. The drainage of the area north of the Port de la Hourquette is subterranean, and the water reappears in the Fontaine d'Ardaing (484/57). The same is seen north of the Pic de l'Homme (486/54) and the Pic de Maubermé, where the water leaves the limestone in the Fontaine de Costeu (485/55) in the valley of the Urets.

On the French side some of the many caves formed by limestone solution have been investigated. A speleological investigation of the Spanish side might well present great difficulties, the amount of water in the caves being considerable and having a very variable level.

The Lago Bargadera, a small lake in the Carboniferous at nearly 2,000 m, loses its water through a sink-hole in its southern shore line where the Maladeta fault separates the Carboniferous from the marmorized Devonian basal limestone. The water reappears in the Fuente de Estua (4°.31'/42°.40') at 1,400 m in the Rio Valarties after circulating through the limestone (Fig. 77-B).

East of the Garona de Ruda, between 1,400 and 1,500 m, several springs occur, together called the Fuentes de Ruda. The water entering the marmorized calcaire métallifère on the southwest slope of the Marimaña granodiorite reappears at these springs. Some of the water of these springs may also derive from the Estany Pudo, a small lake at 2,240 m situated east of the Puerto de la Bonaigua (not exposed on the map), which loses its water in the same Cambro-Ordovician limestone.

The Rio Malo, coming from the Marimaña granodiorite, flows for some distance underground as soon as it enters the calcaire métallifère and reappears at the surface when it reaches other formations.

*) = Well.

In all these instances it has been proved that irrespective of the distance between a sink-hole and the related spring and the intervening topographical feature, both the sink-hole and the spring are situated in the same limestone. The important sink-holes are all situaed on or above the main Tertiary planantion surface (2,000—2,200 m); the springs are all found near the present level of some river. In many cases the sink-holes are situated on fault outcrops where the rock has been shattered. The subterranean water-course need not follow the fault plane as it is strictly bound to the limestone irrespective of its folded shape. Generally the subterranean water-course in crossing a synclinal or anticlinal structure follows the plunge of these

TABLE IX

Thermal springs

number of springs		occurrence	minimum temperature	maximum temperature	average temperature
B. de Luchon northern group	9	5 in granite 4 in schists	39.4° C	63° C	52.7° C
B. de Luchon southern group	20	18 in granite 2 in schists	29.4° C	60.3° C	40.9° C
Lès	1	in granite?	—	—	35° C
Arties	3	in limestone	—	—	40° C
Baños de Tredós	2	in granite	—	—	38° C
Baños de Benasque	5	in limestone and hornfels	31° C	38° C	—

TABLE X

Chemical analysis of sulphurous water of a thermal spring of Bagnères de Luchon (after de Launay, 1899)

	grams/litre
Na_2OCO_2	0.03
NaCl	0.09
Na_2S	0.076
$Na_2OS_2O_2$	0.003
Na_2OSO_3	0.006
K_2OSO_3	0.009
$CaOCO_2$	0.01
$MgOCO_2$	0.0017
Si	0.09
Borium	traces
Phosphate	traces
CO_2	0.014
	0.3297
Residu sec	0.35

structures and maintains a regular gradient without becoming artesian in character.

Thermal springs of varying temperature and more or less of the same chemical composition occur in or near igneous or highly metamorphic rocks (Table IX). Well known are the "Bains de Luchon".

The Baños de Tredós, the Baños de Benasque and probably also the thermal springs of Arties are situated on or near faults. The temperature of these springs is, as records during the last hundred years have shown, slowly decreasing. If they are associated with faulting, this might indicate that these faults are still active.

It has not been possible to obtain chemical analyses of the springs on the Spanish side, but they all are sulphurous. A chemical analysis of sulphurous water from the thermal springs of Luchon is given in Table X.

SUMARIO

Lito-estratigrafía

El establecimiento de la edad de los sedimentos Paleozoicos, ligeramente metamórficos, que se encuentran en la hoya superior del Río Garona, se basa enteramente en la formación característica y muy extendida de "lutita negra" que pudo determinarse como Siluriana en las regiones circundantes por medio de graptolitos. Formaciones de más edad que ésta de características Silurianas, se clasificaron como Cambro-Ordoviciano, y formaciones más recientes como Devónico y Carbonífero.

El Cambro-Ordoviciano aflora en tres grandes unidades estructurales, llamadas el Anticlinal Septentrional y el Domo de Bosost, el Domo de Marimaña y el Anticlinal Central. En la primera y la segunda unidad, los sedimentos están bien diferenciados. Un horizonte espeso de caliza, llamado el "calcaire métallifère", se presenta localmente entre conglomerados irregulares. En general, la formación inferior a este horizonte, es muy arenosa. La parte superior del Cambro-Ordoviciano del Domo de Marimaña se compone de pizarra negra; en el Anticlinal Septentrional, pizarras oscuras alternan con areniscas que tienen un horizonte característico, llamado la caliza "sándwich", por lo que se describe muy bien su apariencia exterior. En el Anticlinal Central no hay "calcaire métallifère".

El Siluriano se compone de sedimentos de grano fino, que se caracterizan por un alto contenido de carbón y azufre. El contenido de hierro es casi normal, el de sílice es bajo. El azufre, que se supone de origen orgánico, se presenta generalmente en la forma de pirita. En la superficie, esta pirita está expuesta a oxidación, por lo que se forma, bajo influencia de la humedad, limonita, la que da el color de óxido de hierro, tan característico del Siluriano. Calizas se encuentran entre los riós Pique y Jueu, ya en la forma de lentes negros, ya en la de nódulos negros. La parte superior del Siluriano se compone a veces de un material de grano más grueso, pero presenta siempre algunas de las propiedades específicas.

El Devónico subdivide en cuatro unidades de sedimentos predominantes.

La caliza basal (D_1) se caracteriza por intercalaciones de pedernal oscuro y areniscas de grano fino a la base, presentándose intercalaciones de pizarras en la parte superior.

Las pizarras de Entecada (D_2), que están bien desarrolladas al norte del Pico de Entecada, se componen de pizarras homogéneas, de color azulado, oscuro hasta negruzco, que cubren la caliza basal. Calizas delgadas están a veces intercaladas, generalmente a la base y en la parte superior. Las pizarras contienen algo de pirita, en cristales grandes o como pigmento. Al oeste del pueblo de Las Bordas, la parte superior contiene areniscas muy delgadas y cuarcitas; más hacia el este, la parte superior está formada de caliza blanca, bien desarrollada, que casi está libre de intercalaciones. Aproximadamente dos Kms. al oeste de la cresta de Las Bordas-Mompius, se encuentran areniscas en la parte superior, aumentando el contenido de arena en dirección este. Las pizarras de color azulado oscuro que afloran al este del Río Areño, difieren en muchos respectos de las pizarras normales de Entecada. Son ligeramente carbonosas y contienen inclusiones de caliza arenosa y detrítica que presenta estratificación cruzada, bien desarrollada.

Las Areniscas de Las Bordas (D_3), que se han desarrollado al este del pueblo del mismo nombre, se componen de cuarcitas, areniscas, areniscas graduadas y pizarras. En la región del Pico de Corbisón, se observa un cambio lateral en la litología. Se encuentran con frecuencia delgadas hojas de arena a la base, que gradualmente se convierten en puras cuarcitas. Cuarcitas muy espesas parecen limitarse a la pendiente oriental del Corbisón.

Las pizarras y calizas de Viella (D_4) se componen de pizarras blandas de color verde, y areniscas delgadas.

El Carbonífero — compuesto principalmente de grauvaces — se encuentra en un sinclinorio de dirección este-oeste; al norte del macizo de la Maladeta se hallaron abundantes fragmentos de plantas — que no pueden ser determinados.

Al nordeste del Pico de Salana, afloran sedimentos de color rojo, de apariencia típicamente Tríasica. Buenos afloramientos son raros, debido a la cobertura de material de morena.

En la orilla occidental del Río Auguamoix, se encuentran depósitos del Mioceno tardío, que se componen de capas alternantes de conglomerados, areniscas, arcilla y lignitos delgados.

Geología estructural

En cada una de las unidades de estructura Herciniana, el tipo de plegamiento es distinto para cada unidad estratigráfica mayor. El Cambro-Ordoviciano se presenta en grandes estructuras de forma de domo, el Devónico en pliegues isoclinales de tamaño medio, el Carbonífero en grandes pliegues concéntricos.

El Siluriano, que actuaba como horizonte móvil, causó el arranque de su cobertura sedimentaria de las estructuras Cambro-Ordovicianas, lo que dió origen a un plegamiento isoclinal del Devónico; estos pliegues isoclinales están generalmente levemente volcados. Las diferencias en plegamiento del Devónico están estrechamente relacionadas con los grandes contrastes en litología entre el Cambro-Ordoviciano y el Devónico.

Las zonas sinclinales, ocupadas por sedimentos Carboníferos y Devónicos, están fuertemente comprimidas. La caliza basal Devónica penetra profundamente en los núcleos sinclinales (Anticlinal de Entecada); en las partes superiores de los sinclinales se presentan muchas complicaciones.

El plegamiento es del tipo de clivaje, que se ha desarrollado en una fase temprana y ha sido precidido probablemente por un débil plegamiento concéntrico y fallas de menor importancia en las unidades competentes.

El clivaje, bien desarrollado en los sedimentos pelíticos, da una delineación en el plano de estratificación de estratos competentes, que corre paralela al plano axial. Las características del clivaje dependen en alto grado de la litología; el desarrollo, las distancias y la desviación (gradual y angular) del clivaje dependen también del tamaño de los granos. Tal vez, a causa de un ligero plegamiento del clivaje primario, se formó un clivaje de fractura a grandes intervalos, que corta el clivaje primario.

En el Cambro-Ordoviciano, se encuentra con frecuencia, en pelitas de grano fino, "knicking" angular de los planos de clivaje en la dirección del buzamiento, pero también se presenta en el Devónico y el Carbonífero. Como no puede esperarse compresión de la roca, "knicked cleavage" da lugar a un acortamiento paralelo al clivaje subvertical. El proceso es el resultado de dilatación, y, por tanto, es un fenómeno del Herciniano tardío, probablemente relacionado con el arqueo de la zona axial.

Intrusión de diques de cuarzo-pórfido — concentrados principalmente en una zona alargada — precedió probablemente a las intrusiones de granito del Herciniano tardío, puesto que los diques presentan amenudo alguna clase de clivaje de fractura.

Una fase tardía de la orogenia Herciniana fué caracterizada por fallas longitudinales. Estas fallas fueron reactivadas durante los movimientos Alpinos, como se ha comprobado para la depresión estructural al norte del macizo de la Maladeta.

Un proceso de mineralización relacionado con la fase de las fallas Hercinianas tardías se produjo en el Anticlinal Septentrional.

Sedimentación

Basándose en su composición, contenido de fósiles y presencia de estructuras sedimentarias, los depósitos arenosos del Paleozoico superior se dividen en tres asociaciones sedimentarias, de acuerdo con Pettijohn:

1. Una asociación de ortocuarcita-caliza, de cuarcitas puras (ortocuarcitas de Las Bordas) y calizas (calizas de Entecada).
2. Una asociación de turbidita compuesta de areniscas homogéneas y graduadas, generalmente laminadas, y lutitas, que presenta estructuras sedimentarias. Una zona en el oeste y sudoeste en que predominan aparentemente areniscas homogéneas, una parte central que presenta estratos continuos y graduados y una predominancia de otras estructuras sedimentarias, y una parte oriental de areniscas homogéneas, falta de estructuras sedimentarias.
3. Una asociación de "molasse" (Carbonífera), compuesta predominantemente de grauvaces, a veces conglomeráticas. Se encuentran frecuentemente estructuras sedimentarias de agua panda y fragmentos de plantas.

La graduación de los estratos de areniscas de grano fino de la asociación de turbidita puede verse a simple vista sólo como un cambio gradual de color, pero se nota muy bien por la curvatura de los planos de clivaje.

El examen de secciones delgadas hace ver que sólo mediante métodos estadísticos puede comprobarse una disminución gradual del tamaño de los granos de un estrato de areniscas graduadas, de abajo hacia arriba. La graduación se expresa también en un aumento gradual de matriz hacia la parte superior de un estrato de arenisca graduada.

De vez en cuando se encuentran en la asociación de turbidita estructuras sedimentarias, amenudo oscurecidas por una tectonización marcada; las direcciones de la laminación cruzada y de la enroscadura enseñan una dirección oriental de transporte.

Las direcciones de corriente que se han medido en la asociación de "molasse", a base de la estratificación cruzada y las marcas de oleaje, varían mucho, limitando, por tanto, su utilidad como indicador de una dirección de transporte.

El surgimiento que se projudo en el oeste durante la época Devónica de Las Bordas (D_3), continuó durante el Carbonífero.

Depósitos del Mioceno tardío, preservados en una zona inestable al norte de la granodiorita de la Maladeta, se componen de una alternancia repetida de conglomerados, arena, arcilla y lignito. La ausencia de componentes graníticos pueden indicar la existencia de un relieve maduro en esta parte del Valle de Arán en una época tardía del Mioceno; grandes zonas ocupadas por las granodioritas estaban, presumiblemente, aún cubiertas de sedimentos.

Morfología

Restos de superficies de aplanamiento se hallan en aproximadamente tres niveles: de 2.400—2.600 m, de 2.000—2.200 m (la más extensiva), y de 1.700—1.800 m. Depósitos del Mioceno tardío se encuentran en una altura de 1.820—1.980 m, inmediatamente debajo de la superficie de aplanamiento de El Pruedo y de la Montañeta que se encuentra en el nivel de 2.000—2.300 m. Es muy defícil la interpretación de una relación entre la superficie de aplanamiento y los depósitos del Mioceno tardío, a causa de la falta de buenos afloramientos. La composición de los sedimentos excluye su formación en una quebrada hondamente erosionada. La presencia de fallas longitudinales puede indicar la preservación de depósitos del Mioceno tardío en una zona de fallas; en tal caso, la edad de la superficie más extensiva de aplanamiento podría ser post-Vindoboniense.

Restos glaciarios, presumiblemente de la última glaciación, se encuentran frecuentemente en las regiones más altas, especialmente en el macizo de la Maladeta. Aquí, la altura del hielo glaciario puede aún distinguirse por las diferencias de alteración, y, en las partes más bajas de la zona estudiada, por la presencia de eráticos. La dirección tomada por los glaciares puede seguirse en muchas partes. Presumiblemente, la fuerza erosiva de los glaciares no fué muy grande.

La erosión post-glaciaria ha cambiado grandemente le topografía glaciaria; la alteración mecánica ha desempeñado un papel saliente. Vastos conos de deyecciones llenaron en parte los valles hondamente erosionados de las regiones más altas. El relieve se ajustó a la litología. Las rocas intrusivas y las cuarcitas forman las rocas más resistentes, depresiones se produjeron, generalmente, en las blandas pizarras Silurianas. Capas sólidas de brechas subrecientes en las pendientes y cimentadas por limonita, se encuentran localmente en zonas cubiertas de despojos Silurianos.

Fenómenos de "karst" se encuentran con frecuencia, tanto en el "calcaire métallifère" como en la caliza basal del Devónico. Dolinas se hallan en, o en un nivel superior a, la superficie más extensiva de aplanamiento, amenudo a lo largo de los linderos entre caliza y otras formaciones, o a lo largo de fallas. El agua que desaparece en las dolinas, puede hallar salida subterránea. Bien conocida es la captación subterránea del curso superior del Río Esera en dos dolinas importantes, el Forat de Aigualluts y el Forat de la Renclusa. Sus aguas reaparecen en el Güells de Jueu, situado en un afluente sur del Río Garona, después de cruzar bajo tierra la divisoria de las aguas *).

*) desviándose así hacia el Atlántico aguas que, normalmente, habrían corrido hacia el Mediterráneo.

Cualesquiera que sean la distancia y el relieve, la dolina y la fuente asociada quedan situadas en la misma caliza. El agua que fluye por las calizas sigue el hundimiento de la esstructura, sin obtener carácter artesiano.

Fuentes termales sulfurosas de diversas temperaturas y más o menos la misma composición química, se encuentran en cinco lugares en, o cerca de, rocas ígneas o altamente metamórficas. Presumiblemente, están relacionadas con las fallas.

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